VOL. 252 JANUARY 1954 No. 1

American Journal of Science

Established in 1818 by Benjamin Silliman

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Published monthly. Eight dollars a year. Seventy-five cents a copy.
Foreign, eight dollars sixty cents. Canada, eight dollars thirty-five cents.

AMERICAN JOURNAL OF SCIENCE

Established in 1818 by Benjamin Silliman

THE FIRST SCIENTIFIC JOURNAL IN THE UNITED STATES

DEVOTED TO THE GEOLOGICAL SCIENCES AND TO RELATED FIELDS

Editor: CHESTER RAY LONGWELL
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Entered as second-class matter at the Post Office at New Haven, Conn., under the Act of March 3, 1879. Published monthly at Sterling Tower, Yale University, New Haven, Conn.

Subscription rate $8 per year (75 cents a number). Student rate $5 per year (application blanks available on request). Postage prepaid to the United States; 35 cents per year to Canada; 60 cents per year elsewhere.

The JOURNAL completed its *first* series of 50 volumes in 1845, its *second* series of 50 volumes in 1870, its *third* series of 50 volumes in 1895, its *fourth* series of 50 volumes in 1920, its *fifth* series of 36 volumes in 1938. Since 1938 the numbering by series has been discontinued, and volumes are annual and bear the whole numbers as of 1818, that for 1939 being volume 237.

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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 35c.)

[American Journal of Science, Vol. 252, January 1954, Pp. 1-25]

American Journal of Science

JANUARY 1954

THERMAL CONDUCTIVITY, CLIMATIC VARIATION, AND HEAT FLOW NEAR CALUMET, MICHIGAN*

FRANCIS BIRCH

ABSTRACT. Measurements of thermal conductivity are reported for some 90 samples of rock from the vicinity of Calumet, Michigan. Mean values, in millicalories per (cm. sec.deg.), are as follows: Jacobsville sandstone, 6.78 ± 0.71; Freda sandstone, 6.81 ± 0.24; Copper Harbor conglomerate, 4.98 ± 0.25; Portage Lake lava series, 5.01 ± 0.11. Some evidence is found for a correlation between conductivity and degree of alteration in the basic lavas. These measurements, and the older determinations of temperature in mines and borings of this region, are discussed with relation to the problem of geothermal evidence for climatic variations during and since the Pleistocene, and to the heat flow. It is concluded that no thermal effect definitely related to climatic change can be discovered with the available data. The local rate of heat conduction to the surface is 0.93 microcalories per (cm^2, sec), with an estimated uncertainty from all causes of about 10 percent.

INTRODUCTION

For a hundred years the temperatures in the mines of the Keweenaw peninsula of Michigan have interested engineers and geologists, but the thermal properties of the rocks have received little attention. Peirce (1903) determined the thermal conductivity of two samples of each of three kinds of rock (trap, amygdaloid, conglomerate); Ingersoll (1932) found the thermal diffusivity for two samples of trap. On the other hand, petrologists distinguish some half-dozen principal types among the lava flows, and there is also variation among the associated conglomerates and other sedimentary formations. The desirability of additional information appeared in the course of an examination of possible effects of climatic variations on underground temperatures (Birch, 1948), of which the Michigan temperatures have been widely cited as an example (Lane, 1923; Hotchkiss and Ingersoll, 1934; Ingersoll, Zobel, and Ingersoll, 1948, p. 119). There is also an opportunity to add to the still meager number of reliable determinations of the rate of conduction of heat to the Earth's surface. New measurements of thermal conductivity of some 90 samples of rock collected in the vicinity of Calumet, Michigan, are presented below; these measurements will be discussed with reference to the two problems of climatic effect and of heat flow.

Darton (1920, p. 58-59) summarized the older measurements of temperature in the mines and borings of this region; the early work by Wheeler, Agassiz, and others suffers from lack of uniformity of procedure and instruments. More recent determinations have been described by Lane (1923), Van Orstrand (1928, 1934, 1935), McElroy (1932), Ingersoll (1931, 1932).

* Paper No. 132 published under the auspices of the Committee on Experimental Geology and Geophysics and the Division of Geological Sciences at Harvard University.

Hotchkiss and Ingersoll (1934), and Fisher, Ingersoll, and Vivian (1934). The best determinations for the present purpose are those by Fisher, Ingersoll, and Vivian in the deeper levels of the Calumet and Hecla Conglomerate mine near Calumet (fig. 1), and by Van Orstrand in a diamond drill hole (Baltic No. 16) of the Copper Range Company, below the 4700-level of the Baltic No. 2 shaft, some fifteen miles southwest of Calumet. In the Conglomerate mine, between depths of 2800 and 5600 feet, the mean gradient is found to be 18.10 ± 0.23 °C./km (9.93 ± 0.13 °F./kft); in the Baltic No. 16 drill hole, between depths of 4800 and 6200 feet, the mean gradient is 16.97 ± 0.12 °C./km (9.31 ± 0.06 °F./kft). The relative uniformity of gradient over this distance along the Keweenaw Peninsula, and in different parts of the stratigraphic section, is evidence of large-scale thermal uniformity, in spite of variations of conductivity from sample to sample that are considerably larger than the difference between these two gradients.

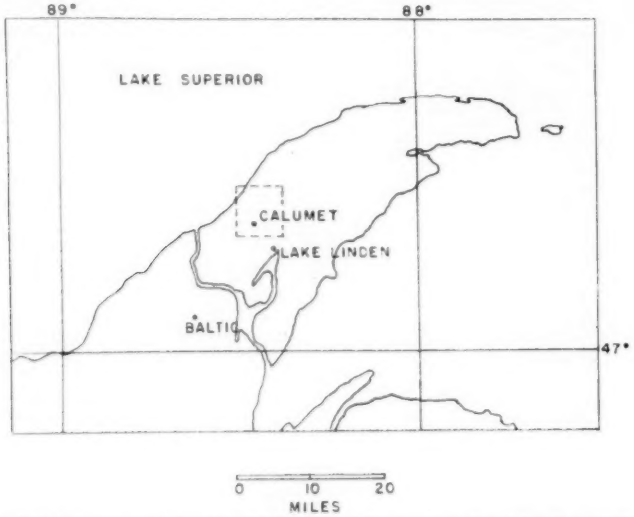

Fig. 1. The Keweenaw peninsula of Michigan. The area enclosed by the small rectangle (broken line) is shown in figure 2.

The geologic structure and petrology of this district have been treated in a number of notable reports: by Lane (1911), Van Hise and Leith (1911), Butler and Burbank (1929), Broderick (1935), Leith, Lund, and Leith (1935), Broderick, Hohl, and Eidemiller (1946), Cornwall (1951), White, Cornwall, and Swanson, (1953). In the last reference, the stratigraphic section for the Calumet area is given as follows, from youngest to oldest:

Jacobsville sandstone, in fault contact with Portage Lake lava series	over 1,500 feet
Freda sandstone	over 5,000 feet
Nonesuch shale	600 feet
Copper Harbor conglomerate	4,300 feet
Portage Lake lava series	over 11,000 feet

For the brief description which follows I am indebted to Dr. White:

"The Portage Lake lava series consists of a succession of about 300 mafic lava flows, with which are interbedded some 20 beds of rhyolite conglomerate which constitute less than 3 percent of the whole formation. The belt underlain by the lava series is bounded on the northwest by outcrops of the younger Copper Harbor conglomerate, and on the southeast by the Keweenaw fault, a major reverse fault along which the lava flows are thrust over the much younger Jacobsville sandstone (Figure 2).

"The Copper Harbor conglomerate (Lane, 1911, pp. 35-39) in the Calumet region is chiefly pebble conglomerate with subordinate sandstone. Most of the pebbles are rhyolitic, with perhaps 10 to 25 percent basaltic or andesitic. This formation contains a few lava flows in the middle and upper parts. The Nonesuch shale is predominantly gray siltstone. The Freda sandstone is brownish or reddish siltstone and fine-grained sandstone, with very minor coarse sandstone and conglomerate.

"The Jacobsville sandstone is not found in contact with the Freda, so that their exact relations are not known. The Jacobsville is mainly fine-grained reddish quartz-rich sandstone, often very friable in outcrop. The rocks are flat-lying except where folded by drag along the Keweenaw fault; the formation may be many thousands of feet thick.

"In the neighborhood of Calumet, the rocks strike northeast and dip northwest toward Lake Superior; dips range from 50° NW in the lowermost part of the Portage Lake lava series to 25° or less in the Freda. In the vicinity of the Baltic mine, 15 miles southwest of Calumet, the dip is about 70° NW. These northwesterly dips reflect the position of the copper district on the south flank of the Lake Superior syncline. Both the Portage Lake lavas and the Copper Harbor conglomerate reappear on Isle Royale with southeasterly dips, on the northern flank of the syncline."

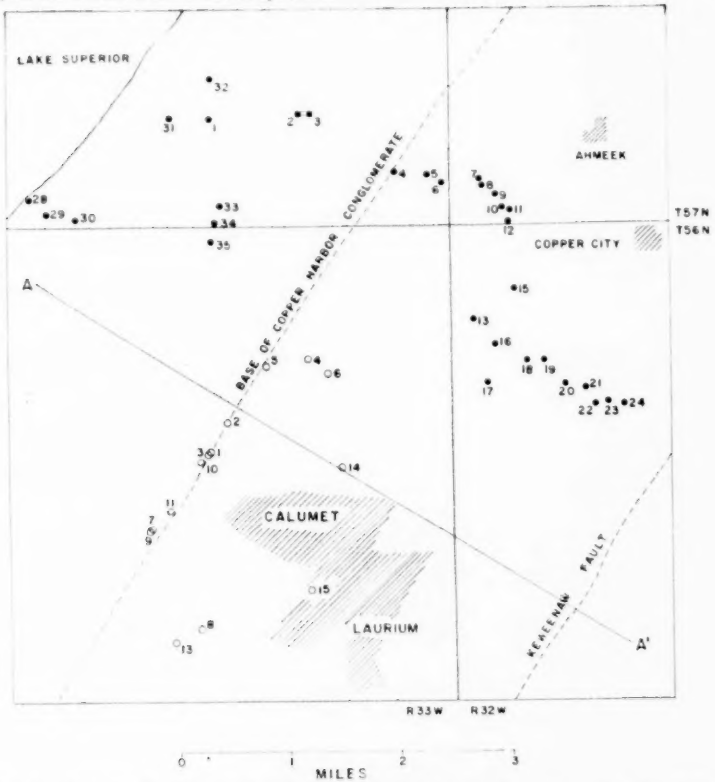

Fig. 2. Vicinity of Calumet. Location of outcrops for numbered samples shown as solid circles; vertical projections of temperature stations in the Calumet and Hecla Conglomerate mine shown as open circles (see table 6).

Figure 3 shows the generalized section (AA′ of fig. 2) in the vicinity of Calumet and also the projection on this plane of the temperature stations of Fisher, Ingersoll, and Vivian in the Conglomerate mine. Since these stations are nearly vertically below the surface contact between the Copper Harbor conglomerate and the top of the Portage Lake lava series, it seemed necessary to include samples of the conglomerate, as it might possess an average conductivity appreciably different from that of the flows. A conglomerate is peculiarly difficult to represent by means of the small specimens used for the measurement of conductivity (1.5-inch disks, 0.25 inch thick), since individual specimens often consist of a single pebble of rhyolite or mafic lava, but an attempt has been made with a collection of some 30 samples.

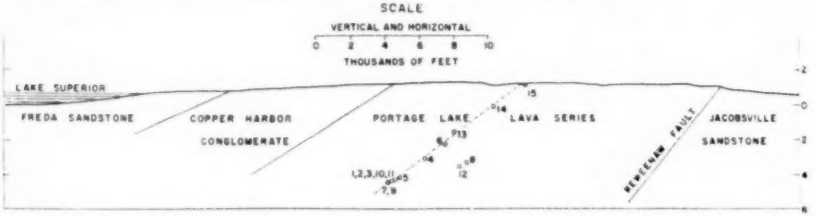

Fig. 3. Simplified geologic section in the vicinity of Calumet (on line AA′ of figure 2). The numbered circles show the horizontal projections on this plane of the temperature stations in the Calumet and Hecla Conglomerate mine.

Besides the samples of conglomerate and sandstone, 27 samples of the flows were collected from the vicinity of Calumet. At the outset, it appeared likely that little variation would be found among the lavas. Butler and Burbank (1929, p. 23) remark: "The great bulk of the flows of the Keweenawan series are basaltic, rather simple mineralogically, and monotonously similar in appearance." It turns out that this similarity conceals differences, particularly of alteration, associated with differences of conductivity, and a much larger number of samples as well as a formidable petrologic program would be necessary to determine with certainty the causes of variation.

When the effect of alteration on conductivity was noticed, it became of interest to examine the amygdaloidal tops of the flows, which are characteristically subject to the most extreme alteration and mineralization, and 10 samples of amygdaloidal flow tops have been studied. The small proportion of amygdaloid, in relation to the whole formation, makes it of relatively little importance for the thermal problems.

THERMAL CONDUCTIVITY

The thermal conductivity was determined by a method described in earlier papers (Clark, 1941; Birch, 1950); in principle, it is a modification of the method of the "divided bar," with samples of pyrex glass, of the same size as the samples of rock, serving as the fixed parts of the bar. Differences of temperature, of a few degrees, across the glass disks and across the sample are measured with differential thermocouples whose junctions are sunk into copper separating disks. A small hydraulic press applies pressure to the flat faces of the sample and glass disks during the measurements. Loads of the order of 5000 psi produce changes of less than 1 percent in the conductivity of most of the basic rocks, but up to 10 percent or even more for some of the porous sandstones. The tabulated figures correspond to a load of 2500 psi.

The mean temperature of the samples was always within a degree or two of 30°C. (86°F.), which is the temperature at about 4700 feet below the surface in the Calumet conglomerate. The variation of conductivity with temperature is known to be small for the basic rocks (Birch and Clark, 1940) and can safely be neglected in the present study. The accuracy of a single determination of conductivity is about 1 percent.

The standard size of sample is a circular disk, 1.5 inches in diameter, 0.250 inch thick; the faces are ground flat and parallel. For some of the samples of conglomerate it was impractical to grind the samples so thin, and the disks were 0.470 inches thick (samples 34a, 35a, 35c); a special calibration was made with a disk of pyrex glass of the same thickness. These carefully ground samples were all weighed and measured, and the densities were found as the ratio of mass to computed volume.

The samples numbered from 1 to 35, with letters denoting individual specimens from the same locality, were collected from correspondingly numbered outcrops in the Ahmeek quadrangle as indicated in figure 2, except sample 26 of the Houghton conglomerate, obtained underground at the Allouez No. 3 shaft, near Ahmeek, samples 14 and 25, from localities in the Laurium quadrangle, just south of figure 2, and sample 27, a piece of building stone from one of several possible quarries in the Jacobsville sandstone. Samples 41a, 41b, 42, and 43 were collected underground. Eleven samples of mineralized amygdaloidal flow tops, numbered 50 to 60, were selected from the Osceola No. 6 shaft dump, the Lasalle No. 5 dump, and from the Kearsarge, Ashbed, and Iroquois lodes. Samples 61 through 65 of the Jacobsville sandstone were obtained from outcrops and quarries in the vicinity of Lake Linden. As indicated by figure 2, the samples of lava and conglomerate are distributed through the stratigraphic column. As nearly as possible, the principal types of lava are represented in numbers roughly proportional to their thicknesses in this section. The textural categories are those which have long been in use in this region (Butler and Burbank, 1929; Cornwall, 1951).

The results of the measurements of conductivity and density are presented in several tables: table 1 for the basic rocks of the Portage Lake lava series, table 2 for the samples of amygdaloidal tops, table 3 for the sedimentary rocks; table 4 is a summary showing mean values. The mean conductivity for all samples of lava, excluding flow tops, is 5.01 ± 0.11 millical/cm.sec.deg., close to values found for other rocks of similar composition (Birch and Clark, 1940). Substantially the same mean is found for the Copper Harbor conglomerate; the dispersion is higher, as one might expect. The samples of the Freda and Nonesuch formations form a fairly close group, with a mean value distinctly higher: 6.81 ± 0.24. This is also the mean value for the few samples of Jacobsville sandstone. As indicated by table 3, the conductivity of many of the sedimentary rocks may be appreciably altered by the compression of a few thousand feet of rock; the higher values corresponding to a simple compressive load of 2500 psi have been used in forming the means. The effect of water saturation has not been investigated for these samples, but it has been shown (Clark, 1941) that in many cases, a moderate compression is nearly equivalent to saturation.

TABLE 1

Conductivity of Basic Lavas

Sample Number	*Description*	*Density*	*Conductivity millical/cm.sec.deg.*
1	Melaphyre from flow in Copper Harbor conglomerate	2.76	5.0
3	Melaphyre from flow in Copper Harbor conglomerate	2.80	4.8
5	Feldspathic melaphyre	2.83	4.3
6	Fine-grained melaphyre, Ashbed Flow	2.93	5.1
7	Glomeroporphyrite	2.77	5.5
8	Feldspathic melaphyre	2.85	4.8
9	Pewabic glomeroporphyrite	2.82	5.4
10A	Melaphyre, Greenstone flow	2.80	4.7
10B	Ophite, Greenstone flow	2.83	4.1
11	Feldspathic melaphyre	2.82	4.4
12	Ophite, Mandan flow	no measurements	
13	Ophite, flow overlying Calumet and Hecla conglomerate	2.89	4.4
14A	Ophite | Laurium quadrangle,	2.85	5.2
14B	Ophite ∫ flow underlying Calumet and Hecla conglomerate	2.84	4.8
15	Ophite	2.89	4.8
16	Ophite	2.86	5.2
17	Ophite, feldspathic	2.82	4.5
18	Porphyritic ophite, Kearsarge flow	2.88	4.7
19	Glomeroporphyrite	2.84	5.6
20	Ophite	2.93	4.5
21	Ophite, Scales Creek flow	2.92	4.6
22	Glomeroporphyrite	2.83	5.6
23	Glomeroporphyrite	2.89	6.6
24A	Feldspathic ophite, Copper City flow	2.90	5.1
24B	Feldspathic ophite, Copper City flow	2.92	4.9
41A and B	Fine-grained ophite, flow overlying Calumet and Hecla conglomerate, Allouez No. 3 shaft, 6 level crosscut	2.87	5.0
42	Ophite, Kearsarge flow, 48 level, Centennial Mine	2.83	6.2
43	Porphyritic trap, Kearsarge flow, 48 level, Centennial Mine	2.86	5.5

TABLE 2

Conductivity of Amygdaloidal Flow Tops

Sample Number	*Density*	*Conductivity*
50	2.87	7.0
51	2.66	5.6
52	2.84	5.6
54	2.77	5.6
55	3.09	6.4
56	3.19	6.6
57	2.84	6.6
58	2.99	9.0
59	2.84	6.1
60	2.60	5.5
Mean of 9, excluding 58		6.1
Mean of 10		6.4

Description of Amygdaloidal Samples

Sample Number	*Place of origin*	*Description*
50	Kearsarge lode (Wolverine No. 2 shaft dump)	Moderately oxidized, highly amygdular, highly fragmental, with about 15%, mostly carbonate, filling.
51	Prospect on Ashbed lode, Sec. 10, T 57 N, R 32 W.	Black, fragmental, with dark red interstitial detritus.
52	Osceola lode (Osceola No. 6 shaft dump)	Slightly oxidized chloritic and chlorite-calcite cellular amygdaloid, sparsely amygdular.
54	Osceola lode (Osceola No. 6 shaft dump)	Moderately well-oxidized, highly fragmental, sparsely amygdular, 15-20% interstitial calcite.
55	Osceola lode (Osceola No. 6 shaft dump)	Gray, pumpellyitized, cellular, with epidote and minor calcite and copper in amygdules.
56	Kearsarge lode (Lasalle No. 5 shaft dump)	Epidotized, cellular, sparsely amygdular.
57	Kearsarge lode (Lasalle No. 5 shaft dump)	Epidotized, pumpellyitized, cellular, quartz and minor epidote fillings. Partly replaced locally by large skeletal calcite crystals.
58	Kearsarge lode (Lasalle No. 5 shaft dump)	Hard, moderately oxidized, cellular, calcite-prehnite-quartz fillings.
59	Kearsarge lode (Lasalle No. 5 shaft dump)	Highly oxidized, cellular, calcite-epidote-chlorite filling.
60	Iroquois lode, Iroquois mine	Moderately oxidized, fragmental, quartz-prehnite filling.

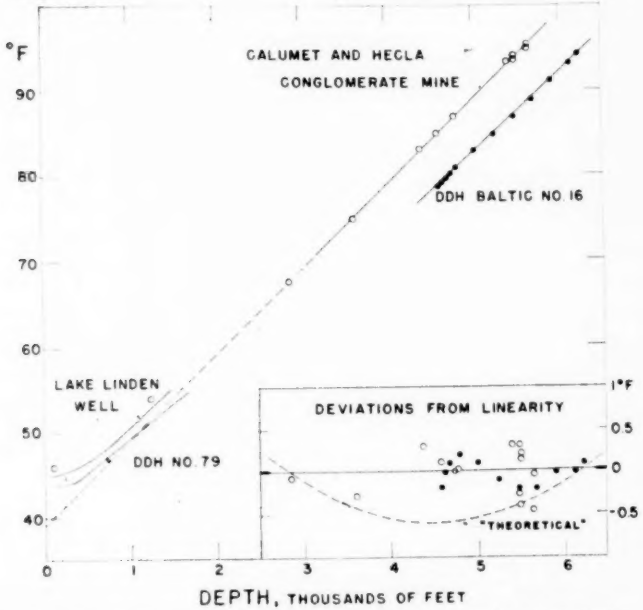

Fig. 4. Temperature versus depth in the Keweenaw peninsula. Open circles, measurements by Fisher, Ingersoll, and Vivian in the Calumet and Hecla Conglomerate mine. Solid circles, measurements by Van Orstrand in Baltic No. 16 drill hole. Measurements by Van Orstrand in the Lake Linden well and in diamond drill hole No. 79 (Copper Range Company) shown as smooth curves. Deviations from linearity shown at lower right hand corner on enlarged scale for points plotted above.

TABLE 3

Conductivity of Sedimentary Rocks

Sample Number	*Formation*	*Density*	*Conductivity K at 2500 psi. millical/cm. sec. deg.*	*Change of K for 2500 psi %*
28	Freda sandstone	2.54	5.3	18
		2.49	5.7	17
29	" "	2.45	7.1	
		2.61	7.4	14
30A	" "	2.54	7.3	
		2.54	6.5	12
30B	" "	2.58	7.3	
		2.57	8.5	9
31A	Nonesuch "shale"	2.52	6.4	
		2.52	7.0	10
31B	" "	2.54	6.2	
		2.53	7.0	13
32A	Copper Harbor conglomerate	2.23	2.9	
		2.21	2.9	4
32C	" " "	2.39	3.9	
		2.39	3.9	8
32E	" " "	2.58	7.1	7
		2.51	2.2	5
32F	" " "	2.38	4.1	6
		2.40	4.1	13
33A	" " "	2.31	5.6	6
		2.32	5.9	4
33B	" " "	2.54	5.3	10
			4.8	22
33C	" " "	2.27	5.1	9
		2.30	5.5	11
33D	" " "	2.41	6.1	7
		2.41	6.2	7
33F	" " "	2.63	6.0	10
		2.63	5.7	
34A	" " "	2.54	4.7	
34B	" " "	2.64	7.9	
		2.68	6.2	
34D	" " "	2.42	3.9	1
		2.53	4.7	7
35A	" " "	2.14	3.4	
35B	" " "	2.36	4.4	10
		2.37	4.2	17
35C	" " "	2.22	3.4	
35D	" " "	2.36	4.2	14
2	" " "	2.65	7.4	1
4	" " "	2.37	4.9	
26	Houghton conglomerate	2.68	7.9	
25	Jacobsville sandstone	2.31	6.3	
27	" "	2.03	5.1	
61	" "	2.17	6.4	8
62	" "	2.41	10.2	6
63	" "	2.14	9.7	8
64	" "	2.02	5.6	
		2.04	5.4	5
65	" "	2.08	5.5	8

TABLE 4

Summary of Measurements of Conductivity

Formation and type	*Number of samples*	*Thermal conductivity millical/cm.sec.deg.* mean	σ	*Mean density*
Jacobsville sandstone	8**	6.78 ± 0.71	1.89	2.15
Freda and Nonesuch sandstone and siltstone	12	6.81 ± 0.24	0.82	2.54
Copper Harbor conglomerate	31	4.98 ± 0.25	1.34	2.42
Portage Lake lava series	37			
Ophites	14	4.85 ± 0.13	0.48	2.88
Glomeroporphyrites	5	5.74 ± 0.20	0.44	2.83
Melaphyre, etc.	7	4.73 ± 0.10	0.27	2.83
Amygdaloidal tops	10	6.4		
All lavas, except amygdaloidal tops	27*	5.01 ± 0.11	0.56	2.85

σ = standard deviation.

* Includes sample 13, omitted from other averages.

** Includes two samples from one hand-specimen.

It is interesting to apply a few simple statistical tests to these measurements. Let us first inquire whether the distribution of conductivity for the lavas as a group differs significantly from a normal (Gaussian) distribution. We apply the chi-square test (Hoel, 1947, p. 186) with six intervals, as follows:

Intervals:	< 4.45	4.45–4.73	4.73–5.01	5.01–5.29	5.29–5.57	> 5.57
No. observed	4	5	7	4	3	4
No. expected	4.3	4.0	5.2	5.2	4	4.3
$(O-e)^2$	0.09	1.00	3.24	1.44	1.00	0.09
$(O-e)^2 \div e$	0.02	0.25	0.62	0.18	0.25	0.02

Then $\chi^2 = \Sigma\,(O-e)^2 \div e = 1.34$, on 3 degrees of freedom (d.f.). This is not significant: there is no indication of a difference from a normal distribution, but the number of samples is too small for a completely satisfactory test.

There is, however, an indication of a difference between the glomeroporphyrites and the ophites. As there are only five samples of glomeroporphyrite, we use the "Student's t" test for the difference between two means (Hoel, 1947, p. 145): on the assumption of equal mean values for these two types, we find $t = 3.45$ for 17 d.f., for which the probability is less than 1 percent. It is thus highly probable that there is a real difference, but it may not be so great as the difference of the sample means of table 4. The "95 percent confidence limits" for the difference between the means are 0.35 and 1.43; that is, the probability that the means differ by less than 0.35 or more than 1.43 is 5 percent. These results depend upon the assumption of normal distributions for both types.

The 8 samples of Jacobsville sandstone show a large dispersion; the Student's t test leads to 6.78 ± 1.68 for the 95 percent confidence limits for the mean. Even these wide limits of statistical uncertainty probably do not represent the real uncertainty since the collection of samples is inherently

biased in favor of the harder, mechanically stronger, rocks. In a formation like the Jacobsville, containing much friable, practically uncollectible material, the outcrops tend to represent the more resistant portions. Furthermore, since the Jacobsville formation is relatively flat-lying, there is no opportunity to sample it in depth, as could be done for the dipping lavas and conglomerates, northwest of the Keweenaw fault. The Jacobsville sandstone is quartz-rich, and when compact shows high conductivity (samples 62, 63); the differences are due primarily to differences of porosity, though grain size may have some effect. If the formation contains a higher proportion of rocks of high porosity and low conductivity than does the small sample which has been studied, then the average conductivity may not exceed 5 or so, in spite of the statistical "confidence limits."

The difference between the ophites and the glomeroporphyrites, for which there is no obvious explanation, may be related to alteration. The principal primary minerals in the rocks of both groups are feldspars (labradorite or andesine), pyroxene, olivine, and, in some cases, magnetite. The conductivity of the plagioclase feldspars does not vary greatly with composition and may be taken as about 4.2 for these rocks (the unit of millical/cm. sec.deg. will be used for conductivity without further specification). Fresh olivine and pyroxene have a conductivity of about 10. Fresh gabbros and diabases are usually found to have conductivities in the expected range, between about 4.5 and 5.5, with the denser rocks showing the higher conductivity. In the Michigan lavas, however, the denser lavas (aside from the amygdaloids) tend to have the *lower* conductivities, and it was at first assumed (a) that alteration would lower the conductivity, and (b) that alteration was most severe among the ophites. Examination of thin sections of some of the rocks showing the most extreme values soon indicated the reverse: the highest conductivities were shown by the most intensely altered rocks.

In an effort to account for this unexpected result, a small collection was gathered from the Harvard Mineralogical Museum of minerals common among the alteration products, such as chlorite, serpentine, and hematite. These samples, all massive aggregates, were studied in the usual way, and the results are given in table 5.

TABLE 5

Conductivity of some Monomineralic Aggregates

Sample	*Density*	*Conductivity*
Serpentine A	2.65	5.9
Serpentine B	2.41	4.3
Talc	2.79	7.1
Chlorite	2.79	12.5
Hematite	4.1	25
Magnetite	4.6	12.5

These values serve to illustrate the possible effect of alteration in these rocks. We have to consider chiefly the alteration of olivine to serpentine, chlorite, magnetite, and hematite, and of magnetite to hematite (Butler and Burbank, 1929, p. 34; Broderick, 1935, p. 517; Cornwall, 1951, p. 162, 175). In addition, the plagioclase, usually least altered, may be "slightly to mod-

erately chloritized," and the pyroxene "slightly altered to chlorite and hematite" (Cornwall, p. 162). It seems possible that the general tendency of such alteration is to increase the conductivity. Since it is probable that the glomeroporphyrites are typically more altered than either the ophites or melaphyres, alteration may account for the apparent correlation between relatively high conductivity and glomeroporphyritic texture. Only a close correlation of the actual mineralogy with conductivity would enable us to be sure whether the differences can be accounted for in detail in this way.

Incidentally, these considerations may account for the relatively high values of conductivity found by Bullard (1939) and recently confirmed by Mossop and Gafner (1951) for the Ventersdorp lava. Aside from the amygdaloidal lava, for which the values of 7.7 to 8.6 may be attributed to the inclusions of quartz, calcite, and chlorite, the non-amygdaloidal samples give values of 7.2: this is considerably higher than seems likely for an unaltered andesitic lava, and the difference is perhaps attributable to extreme alteration.

The new measurements of conductivity lead to a mean thermal diffusivity for the lavas of about 0.009 $cm^2/sec.$ at 30°C., slightly higher than the values found by Ingersoll, 0.0077 and 0.0075, which correspond to conductivities of about 4.3 to 4.4. Several of our samples fall in this range.

Peirce (1903, p. 658) gives the following values of conductivity for rocks from the Calumet region: trap, 3.1, 3.6; conglomerate, 4.7, 5.2; amygdaloid, 3.5, 3.4: like many of the older measurements of thermal conductivity of rocks, these differ appreciably, and unaccountably, from more recent values.

TEMPERATURES AND GRADIENTS

The measurements of greatest value for the determination of undisturbed gradient, and subsequently of the flow of heat, are the two series of careful observations at deep levels already mentioned. In addition, there are several series of observations by Van Orstrand for relatively shallow wells and diamond drill holes which throw light on the near-surface conditions, and are thus of particular interest in connection with the problem of climatic effect. There are also a number of observations of unknown quality at scattered levels in mines over a distance of fifty miles along the Copper Range. There is no consistent series of measurements in a single region for the whole range of depth through which the mines have operated.

Fisher, Ingersoll, and Vivian have described their technique in several publications (1934; Ingersoll, 1931, 1932); the consistency of their measurements, in carefully selected, actively advancing drifts in the Calumet and Hecla mine, is shown in table 6 and figure 4. These measurements seem not to have been published, except as part of a curve given by Hotchkiss and Ingersoll (1934, p. 117); they have been made available through the courtesy of the Engineering Department of the Calumet and Hecla Company, together with the pertinent comments on location and other circumstances. In addition to the original stations numbered 1 through 8, three others of nearly equal value, numbered 9 through 11, are included; these are all in drill holes as indicated. No. 12 is at the end of a diamond drill hole of unstated depth, drilled from a drift on the Kearsarge lode and originally measured in 1921;

a measurement ten years later showed a difference of only one-half degree. Stations 13, 14, and 15 are in old workings at the end of long drifts; the exact procedure in taking the temperatures is not recorded, but it is surmised that the thermometers were placed on the floor in a corner of the drift rather than in drill holes, as for the other stations. No. 13 is on the 49 level of Hecla No. 12 shaft, a downcast shaft; No. 15 is on the 1 level of No. 14 shaft (Osceola lode), an upcast shaft. These circumstances of air circulation may introduce unknown disturbances. In any case, these three determinations are of lower value than the others; as Misener and his colleagues have shown (Misener, 1949), the effect of ventilation in old workings may penetrate as much as 20 feet or more into the rock. No fewer than 8 of these stations are between depths of 5400 and 5700 feet, and the shallowest of the best determinations is at 3600 feet. On the other hand, the deepest stations are scattered horizontally over a distance of two miles, with little horizontal variation of temperature.

For the determination of the gradient at the lower levels, a least-squares straight line has been fitted to the 12 best determinations (1 through 12). The slope of this line is 18.10 ± 0.23°C./km (9.93 ± 0.13°F./kft). The deviations are shown in the lower righthand corner of figure 4; the greatest deviation is several tenths of a degree, which is comparable with the uncertainty of the measurements as established by repeated readings. There seems to be no significant departure from linearity, the largest deviations occurring at depths at which other observations show little or no deviation. The least-squares slope is slightly higher than the value given by Fisher, Vivian, and Ingersoll (9.7°F./kft). Station 13, which was not used for the determination of this line, departs from it by only 0.05°C., so that we may say that there is no evidence for any systematic departure from linearity between 2800 and 5600 feet.

Measurements by Van Orstrand were taken with mercury maximum thermometers at 12 depths in a single diamond-drill hole in the Baltic Mine of the Copper Range Company. The drill hole starts at the 4700 level of the Baltic No. 2 shaft, has a dip of about 64° toward S27°E, and has an inclined length of 1847 feet; the deepest observation is 6228 feet vertically below the surface of the ground. This hole cuts across the flows, and the bottom must be close to the Keweenaw fault. A small flow of water from the hole, "about ½ barrel per day," was recorded. As shown by table 7, the consistency of these measurements is also extremely good; the standard deviation of a single point from the least-squares straight line is 0.13°F. or 0.07°C. The slope is 16.97 ± 0.12°C./km (9.31 ± 0.06°F./kft). The temperatures are 6 to 7 degrees (F.) lower than those at the same depth at Calumet.

Temperatures were also obtained by Van Orstrand in three shallow diamond drill holes several miles southwest of the Baltic Mine (table 8). The uppermost hundred feet or more is in glacial drift, the lower part in melaphyres (No. 72 and No. 75) or ophites (No. 79). The temperature begins to rise at a depth of about 300 feet, at a rate of 7.8°F./kft for No. 72, 7.7°F./kft for No. 79 (432 feet to 1654 feet). It seems likely that water circulates not only in the overburden but also in the uppermost hundred feet

TABLE 6

Temperatures in the Calumet and Hecla Conglomerate Mine

Station Number	Shaft	Depth,* feet	Temperature, °C. Observed	Calculated**	O − C.	Remarks
15	Calumet 14	115	7.76			Osceola lode; Notes A, B
14	Calumet 5	1,242	12.10			Notes A, B
13	Hecla 12	2,851	19.72			Notes A, B
6	Tamarack Jr.	3,604	23.75 ± 0.10	23.92	−0.17	
4	Tamarack Jr.	4,379	28.35 ± 0.15	28.19	+0.16	
8	Hecla 12	4,578	29.36 ± 0.12	29.29	+0.07	Kearsarge lode
12	Red Jacket	4,780	30.43	30.40	+0.03	Kearsarge lode
5	C	5,408	34.05 ± 0.10	33.87	+0.18	
10	C	5,481	34.44	34.27	+0.17	14-foot hole
11	A	5,482	34.13	34.28	−0.15	Footwall, 34.18; hanging wall, 34.08 14-foot holes
3	C	5,488	34.39 ± 0.05	34.31	+0.08	14-foot hole
1	C	5,490	34.44 ± 0.08	34.32	+0.12	
2	C	5,494	34.11 ± 0.08	34.34	−0.23	
9	A	5,647	34.94	35.19	−0.25	Footwall, 34.79; hanging wall, 35.09 14-foot holes
7	A	5,647	35.17 ± 0.08	35.19	−0.02	

* Measured from surface point vertically above station.

** Calculated from: $T(°C.) = 4.04 + 0.005516\ x$, where x is depth in feet.

Note A: Thermometer on floor?

Note B: Not used for least-squares calculation.

or so of the bedrock. It is known that large flows of water are encountered by workings which break through into the glacial drift, although the lower levels are notably dry (Lane, 1895, p. 769; McElroy, 1932, p. 4; Butler and Burbank, 1929, p. 100). The temperature of the groundwater is evidently close to 44°F., in good agreement with Lane's estimate for the surface temperature of the ground, after a rough allowance for the effect of snow cover (43.7°F., Lane, 1923, p. 709). Where groundwater can circulate, however, this temperature is brought to deeper levels, as shown by these drill holes, where the temperature is still about 44°F. at depths of 300 feet.

There is another series of measurements by Van Orstrand in the Calumet and Hecla well No. 2, at Lake Linden, near the shore of Torch Lake. This well, 1518 feet deep, was drilled in 1894, and a few observations were made by E. S. Grierson shortly after its completion. Van Orstrand's measurements, in 1930, are within about 1°F. of the older ones. There appears to have been no artesian flow; the water is reported as standing at a depth of 20 feet. The diameter was 8 inches, except for 68 feet of 12-inch and 10-inch pipe at the top. The temperature near the top is close to 45°F.; it rises with depth at an increasing rate to a depth of 900 feet, and thereafter shows a nearly uniform rate of increase of 9.6°F./kft. This well is in the Jacobsville sandstone. It is noteworthy that the gradient in the deeper part is close to the values found in the lava flows at the greater depths, and that the near-surface temperatures are close to those found in the diamond drill holes 18 miles away.

Van Orstrand has published mean gradients for various ranges of depths, based on a considerable number of measurements in various mines of this district. These average values do not apply to a particular locality, and they must be affected in an unpredictable way by the circumstance that different depths were accessible in different mines. The results, however, are not greatly

TABLE 7

Temperatures in Diamond Drill Hole Baltic No. 16

(Baltic No. 2 shaft of Copper Range Company. Measurements by C. E. Van Orstrand).

	Temperature, °F.		
*Depth, feet**	*Observed*	*Calculated*	*O − C.*
4590	78.5	78.7	−0.2
4635	79.1	79.1	0
4680	79.6	79.5	+0.1
4736	80.1	80.1	0
4793	80.8	80.6	+0.2
5017	82.8	82.7	+0.1
5242	84.7	84.8	−0.1
5467	86.7	86.9	−0.2
5691	88.8	89.0	−0.2
5916	91.1	91.1	0
6141	93.1	93.1	0
6228	94.1	94.0	+0.1

Least-squares solution: $T = 36.0°F. + 9.31 \pm 0.06$ °F./kft.
$= 2.2°C. + 16.97 \pm 0.12$ °C./km.
$\sigma = 0.13°F. = 0.07°$ C.

*Below a surface point vertically above the collar of the drill hole: elevation above sea level 1230 feet.

TABLE 8

Temperatures in Diamond Drill Holes of the Copper Range Company

(After Van Orstrand)

Diamond drill hole	*Vertical depth Feet*	*Temperature °F.*	*Remarks*
No. 72	106	43.9	Water between
Completed Dec. 4, 1929	212	44.0	70 and 75 feet.
Temperatures logged Sept. 1930	318	44.4	Overburden of
	425	45.2	glacial drift
	531	46.1	to about 190
	638	46.9	feet.
No. 75	22	45.0	Overburden to
Completed Feb. 11, 1930	67	44.0	110 feet.
Temperatures logged Sept. 1930	111	43.9	
	223	44.0	
	334	43.9	
No. 79	87	43.9	Overburden to
Completed July 30, 1930	173	43.9	134 feet.
Temperatures logged August 1931	260	44.0	
	346	44.3	Water standing
	432	44.9	at depth of
	518	45.3	about 85 feet.
	604	46.0	
	690	46.7	
	820	47.9	
	952	48.8	
	1083	49.9	
	1215	51.0	
	1347	52.0	
	1479	53.1	
	1654	54.3	

TABLE 9

Temperatures in Lake Linden Well

(Calumet and Hecla Artesian Well No. 2, Lake Linden, Completed Dec. 22, 1894)

	Vertical Depth Feet	*Temperature °F.*
Temperatures by E. S. Grierson, 1894-1895	100	45.5
	200	45.5
	500	47.5
	1000	51.5
	1494	55.5
Temperatures by C. E. Van Orstrand, September, 1930	100	44.9
	250	45.5
Water standing at depth of 20 feet	375	45.9
	500	46.6
	625	47.3
	750	48.3
	875	49.2
	1000	50.4
	1125	51.6
	1250	52.8
	1375	53.9
	1435	54.6

different from the gradients found above; for example, the mean gradient between 2000 and 5370 feet, for "70 observations in mines," is given as 16.09°C. km (H. C. Spicer, in "Handbook of Physical Constants," Birch, Schairer, and Spicer, p. 282); the mean from 100 feet to 5370 feet is given in the same reference (see also Van Orstrand in Gutenberg, editor, 1951, p. 117) as 15.53°C. km. The lower mean to shallow depth is the result of water circulation, with nearly uniform temperature, in the uppermost few hundred feet.

Figure 5, which shows the temperatures above the depth of 3000 feet, includes the measurements by Agassiz, as given in a compilation in the files of the U. S. Geological Survey prepared by Van Orstrand. These are revised figures which do not include the erroneous determination at 105 feet which was the basis for Agassiz' preliminary estimate of a mean gradient to 4580 feet of only 4.5°F./kft (8°C./km) (Agassiz, 1895; see also Rept. British Assoc. Adv. Sci. 1901, p. 64-66; Lane, 1895, p. 773[1].) It will be necessary to refer to these when we consider the evidence for a glacial effect.

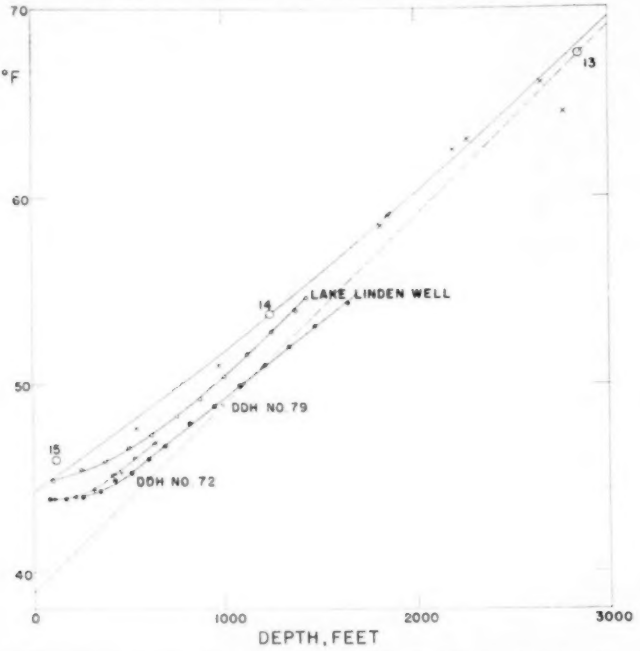

Fig. 5. Temperature versus depth at shallow levels. For the numbered points, see table 6; crosses (x) by Agassiz; other determinations by Van Orstrand. Broken line is least-squares straight line for deeper points in the Calumet and Hecla Conglomerate mine.

[1]As this reference may not be readily available, Lane's characteristic comment on these measurements is reproduced:

"President Alexander Agassiz has just given the following temperatures: At 105 feet, 59°F.; at 4,580 feet, 79°F., that is, 1°F. for 223.7 feet. Since, however, at 105 feet the rock temperature should be near the mean annual temperature of the locality, and since the mean annual temperature of Calumet is, according to all isothermal maps, near 39°, and a mean annual temperature of 59° is found somewhere near Tennessee, I do not think we can safely assume a gradient very much less than 1° in 100 feet, after all."

DISCUSSION

The available evidence with regard to temperatures and conductivities will now be examined with reference to two major topics: (1) the possibility of detecting an influence of climatic variation, and (2) the computation of the conducted flow of heat. The first depends upon the discovery of significant departures from linearity of the temperature-depth curve not adequately accounted for by known disturbances, especially at shallow depths; for the second, we need only the relatively well-established average gradients in the deeper levels. We begin by estimating some effects of the topography.

Topographic correction.—In the neighborhood of Calumet, the Keweenaw peninsula reaches an elevation of about 1200 feet, with little nearby relief, and under most circumstances one would not trouble to compute a topographic correction. The proximity of Lake Superior, however, with its surface at about 600 feet above sea level, and a depth of some 600 feet in this region, long ago led Wheeler (1886) to suggest a relation between the low temperatures in the mines and the presence of this body of cold water. This proposal was rejected by Lane (1923) on the basis of an appeal to common sense and reference to calculations by B. O. Peirce which do not apply directly to this subject. An approximate calculation of the effect of the Lake is readily performed on the assumption that the problem is, closely enough, two dimensional[2].

The basin of Lake Superior contains water at about 39°F. and a few miles off shore reaches a depth some 1200 feet below the surface of the ground at Calumet; at the same level in the mines, the temperature is about 51°F. On a horizontal plane at a depth of 1200 feet (or sea level) the temperature thus varies from 51°F. near Calumet to 39°F. at the bottom of Lake Superior. Let us find the effect of this departure from a uniform temperature of 51°F., at a point 6000 feet below the outcrop of the base of the Great Conglomerate, near Calumet. The steady state disturbance Δv, at a point P, due to the variation of temperature ΔT over a plane surface, is given (Birch, 1950, Eq. 133) by $\Delta v = \pi^{-1} \int_{-\pi/2}^{\pi/2} \Delta T \, d\Psi$, where $d\Psi$ is the element of plane angle at P subtended by a strip on which the disturbance of temperature is ΔT. This integral is approximated as a sum of terms corresponding to finite increments of angle, say θ_i; expressing these increments in degrees, we obtain $\Delta v = \Sigma_i \, \Delta T_i \, \theta_i / 180$, where ΔT_i is the mean disturbance of temperature for the angular increment θ_i.

The cross section at Calumet is approximately as shown in figure 3, and we suppose that this section persists indefinitely in the direction normal to itself. The lake shore is reached at about 3 miles to the northwest. The lake bottom gradually descends about 600 feet in the next few miles, and thereafter remains close to this level for nearly 50 miles; we suppose that it continues indefinitely at this level. The lake bottom subtends an angle at our

[2]The three-dimensional problem can, of course, also be treated by the same method; but since the two-dimensional correction is small there is little to be gained by the more laborious calculation.

6000-foot station of approximately 15°. The disturbance produced by the lake is consequently $-12° \times 15 / 180$, or $-1°$F., at this point. The effect of the rest of the topography is shown in table 10; the mean departure from the normal temperature of 51°F. at a depth of 1200 feet is estimated for 15° increments. The total disturbance is about 2°F., of which the lake contributes one-half. Thus the effect is small but perceptible. The disturbance of gradient is 2°F. in 6000 feet, or 0.3°F./kft, about 3 percent of the observed value. The cooling effect of Lake Superior first increases as the depth increases, reaching a maximum of around 4°F. at 40,000 feet, and thereafter slowly diminishing.

As Lane recognized (1934, p. 540), the presence of as much as several hundred feet of overburden, in which there is substantially no rise of temperature, means that the surface which is really the datum for underground temperatures is the bedrock surface, or more exactly, the surface at which the circulation of water ceases. Above this surface, the temperature is a uniform 44-45°F., as shown by the measurements in the wells and drill holes. Thus the true relief for thermal corrections is different from, and possibly somewhat greater than, the topographic relief. The consequences of this for depths of more than several thousand feet are probably negligible; as we have just seen, even Lake Superior produces an effect of only 1°F. at 6000 feet. The effect of the irregularity of the depth of active circulation will be greatest at the shallow stations, where the depth in which conduction takes place will be uncertain by an appreciable fraction of the whole depth. But at any depth it is difficult to make this correction to a precision of 10 feet, and this corresponds to a temperature difference of 0.1°F., which is of the same order as the deviations from linearity.

TABLE 10

Topographic Correction at 6000-foot Depth, near Calumet

Angular increment degrees		*Surface elevation above sea level, feet*	ΔT. °F.
NW			
0 – 30		1200	0
30 – 45		1100	–1
45 – 60		1000	–2
60 – 75		800	–4
75 – 90	(Lake Bottom)	0	–12
SE			
0 – 75		1200	0
75 – 90		800	–6
			–25

$$\Delta v = -25 \times 15 / 180 = -2°\text{F.}$$

The effect of Lake Superior will be slightly increased by the relatively high conductivity of the Freda formation, which was not taken into account in the foregoing estimate. The ratio of conductivities, Freda/lavas, is 1.36. A rough computation gives for the additional disturbance at the 6000-foot point no more than 0.5°F.; this will be disregarded.

The topographic correction has also been computed for the top and bottom of the Baltic No. 16 drill hole, again on the assumption of two-dimen-

sional symmetry. In this region, the lake is no longer important, but the ridge which is defined by the outcrop of the lavas falls away sharply on both sides, at about one-half mile from the drill hole to the southwest, at about two miles or so to the northwest. The results are that the effective depth for the top of the hole is 90 feet less, and for the bottom of the hole 140 feet less than the depths shown in table 7 which are simply the vertical distances below the elevation of 1230 feet. Thus the temperatures at these two points are respectively 0.9°F. and 1.4°F. lower than they would be if the surface of the ground were a horizontal plane at this elevation. The correction to the gradient is 0.5°F./1640 feet, (0.3°/kft), or about 3 percent, as at Calumet.

Variation of conductivity.—The laboratory measurements of conductivity show appreciable variations among the individual samples which have already been discussed: the range for the lavas (excluding the tops of flows) is from 4.1 to 6.6, for the conglomerates, from 2.2 to 7.9. On the other hand, the measurements of temperature may be used to show that the average conductivity for intervals of depth of some hundreds of feet is relatively uniform.

The standard deviation of a single point from the least-squares straight line for the observations below 2800 feet in the Calumet and Hecla mine is 0.15°C.; this is consistent with the "estimated probable errors" shown in table 6, so that nearly all of the observations are as close to the best line as can be expected. Let us suppose, however, that the deviations result from variations of conductivity, instead of from instrumental errors, topographic effects, and local disturbances. Then we may form the mean gradients and mean conductivities shown in table 11. The conductivities are calculated on the assumption that the mean gradient, 18.10°C./km, corresponds to a mean conductivity of 5.01 millical/cm.sec.deg, as found for the lavas, and that the

TABLE 11

Mean Gradients and Estimated Conductivities for Intervals of Depth

Station	*Depth, feet*	*T, °C.*	*Means for intervals* *Gradient °C./km*	*Conductivity millical/cm.sec.deg.*
	400	(7.22)		
13	2,851	19.72	(16.7 ± 1.3)	(5.43)
6	3,604	23.75	17.7 ± 0.6	5.12
4	4,379	28.35	19.5 ± 0.6	4.85
12	4,780	30.43	17.0 ± 1.2	5.33
3	5,488	34.39	18.3 ± 0.7	4.96
Means for whole depth			18.10 ± 0.23	5.01 ± 0.11

heat flow, gradient times conductivity, remains the same for all intervals. The conductivities so computed are all within the range of the measured values; the hypothesis that the deviations arise from variations of conductivity is not notably inconsistent with the measurements of conductivity. When, however, we consider the errors of the observations, it is clear that there is little or no real evidence that the large-scale conductivity departs appreciably from uniformity.

Table 11 includes an estimate for the interval of depth, 400-2851 feet, based on a temperature at 400 feet of 45°F. or 7.22°C. as given by the measurements in the shallow diamond drill holes. The corresponding value of

conductivity is 5.43, about 8 percent higher than the average conductivity of the lavas but within the possible range. If the vertical heat flow at the Baltic No. 16 drill hole is equal to that at Calumet, the mean conductivities for the intervals of temperature measurements must be in the ratio of 18.10/16.97, or 1.066/1, which is also an acceptable ratio, leading to a conductivity of 5.34. The mean gradient in the upper part of the section at Baltic, between 400 feet and 4590 feet, is 14.9°C./km; this would require a mean conductivity, again on the assumption of heat flow as at Calumet, of 6.1. This value, near the upper limit of the range found for the Calumet samples, suggests that the rocks in the Baltic region are better conductors than those at Calumet by some 10 to 20 percent.

Although the conductivity in the deep section of the Calumet and Hecla mine appears to be substantially uniform, it does not follow that the correct mean value is the mean for our samples. The distribution of samples among the major textural types roughly corresponds to the distribution for the whole series in the vicinity of Calumet, but a relatively small number of the many individual flows are represented, even by a single sample, and the correlation between conductivity and textural type is still inadequately established. It has also been assumed that no systematic change of conductivity occurs on passing down the dip. Nevertheless, since the standard error of the mean conductivity of the lavas is only 2 percent, and of the mean for the Copper Harbor conglomerate about 5 percent, it seems that a conservative estimate of error of the adopted mean need not exceed 10 percent.

Climatic variation.—The hope of Lane (1923) and of Hotchkiss and Ingersoll (1934) that the chronology of postglacial climatic variations might be discovered through analysis of the temperature-depth relations seems to have been the principal motivation for the careful measurements in the Calumet and Hecla mine in the 1930's. By this time, however, this mine was in active operation only at the deeper levels, and for the uppermost few thousand feet Hotchkiss and Ingersoll were dependent upon the measurements of Agassiz, which are known to be unreliable at the deeper levels; for example, the difference between Agassiz' determination at 4580 feet and that of Fisher, Ingersoll, and Vivian is 5°F., and differences of several degrees occur at other levels. Consequently, the value of the older measurements at the upper levels is also doubtful (Fisher, Ingersoll, and Vivian, 1934, p. 529). In addition, Hotchkiss and Ingersoll made use of the new measurements at Stations 13, 14, and 15 (table 6), and adopted for the temperature at zero depth a value computed by Lane's method, 44.3°F. (6.83°C.). A smooth curve (the solid line of figure 5) bearing a very reasonable relationship to the somewhat scattered points was then drawn, and the discussion of climatic effect was based on the deviations of this curve from linearity. In this procedure, the smooth curve was given an importance scarcely justified by the amount or accuracy of the data on which it was based; the deviations of many of the observations from the smooth curve are larger than the deviations which are supposed to indicate climatic change.

The variation of temperature with depth in the upper levels, moreover, may differ systematically from the smooth curve of Hotchkiss and Ingersoll.

Van Orstrand's measurements at shallow depth in the well at Lake Linden and in the three diamond drill holes near Baltic all show a common feature, namely, a nearly constant temperature of 44-45°F. to a depth of about 400 feet. It seems very likely that this is brought about by circulation of groundwater. If a similar condition prevails in the neighborhood of Calumet, then the straight line determined by observations below 2800 feet may hold almost to the surface; the linearly projected temperature at 400 feet is 43.2°F., only about 1°F. below the temperature of the groundwater. A curve drawn to connect the straight-line portion between 2800 and 5600 feet with near-surface temperatures such as those in the bore-holes would thus have much less curvature than that of Hotchkiss and Ingersoll, and would, of course, lead to quite different conclusions about the climatic variation. Comparison of these curves shows that there probably is no unique "temperature distribution" at the shallow depths which can be ascribed to climatic effect.

The intercept at zero depth of the straight line determined by the best measurements is not 32°F., as Lane supposed, but 39.3°F., close to the present mean annual air temperature at Calumet. This coincidence is probably without significance, as the surface temperature which is important for the deeper temperatures is the temperature of the groundwater, which, as we have seen, persists with little change to a depth of several hundred feet. The temperature of the groundwater, in turn, is related, not only to the mean annual air temperature, but also to the duration of snow cover, as Lane showed. It is perhaps worth recalling that in some of the earliest measurements in this district, Wheeler (1886; also Darton, 1920) found temperatures at depths of roughly 100 feet of about 42-44°F., notably lower than the later measurements; these points are consistent with a linear change between 400 and 5600 feet, with uniform temperature above 400 feet.

On the other hand, diamond drill holes No. 72 and No. 79 show gradients in their deepest parts of about 14°C./km, whereas in the nearby Baltic No. 16 hole, below 4300 feet, the gradient is 17°C.km. This difference, curiously enough, is approximately the difference to be anticipated on the basis of a schematic chronology of glaciation (Birch, 1948, p. 753)[3], but without further knowledge of conductivity and water movement in this area we cannot conclude that it arises from such a cause. In the well at Lake Linden, a gradient of 17.5°C./km is reached below 875 feet; the curvature for the upper part of the Lake Linden well can evidently not be accounted for in terms of the same climatic variation which would be required for the holes near Baltic.

A climatic effect as large as that of the schematic chronology should be indicated, under favorable conditions, by the measurements between 3000 and 6000 feet, as a regular departure from linearity reaching a maximum deviation for this interval of about 0.6°F. The "theoretical" deviation curve based on this chronology is plotted as a broken curve at the lower righthand corner of figure 4; it bears no evident relationship to the actual deviations from linearity, either for the Calumet and Hecla mine, or for the Baltic No. 16 hole.

[3]The correction for $k = 0.01$ cm^2/sec and a depth of 1 km is 1.83°C., instead of 1.73°C., given in the reference; the value at 1.5 km is 2.08°C.

The numerous accurate measurements by Misener and his collaborators in the mines of Ontario and Quebec show no disturbances which can be definitely assigned to climatic variation. This region is relatively unfavorable for finding an effect of deglaciation, since the surface temperatures are close to the freezing point (Leith, 1952). Many of these surveys, however, show horizontal fluctuations of temperature more than sufficient to mask the climatic effects, even in regions where they might be expected to be greatest.

Radiocarbon dating of various late-Pleistocene events, recently reviewed by Flint and Deevey (1951), has tended to shorten the interval between the present time and the last deglaciation, as well as the interval back to the post-glacial thermal maximum. There seems also to be a tendency to shorten the times required for glacial advances and retreats. Such negative evidence as we can derive from geothermometry supports the view that the durations of the glacial stages, or at least of the Wisconsin ones, have been overestimated in the past.

As matters now stand, we have no example of curvature of a temperature-depth plot which can be traced with certainty to a climatic effect. This does not necessarily mean that the effect does not exist, nor even that it is beyond the limits of detection. It is clear, however, that to find a climatic effect it will be necessary to obtain a larger quantity of consistent observations of temperature, particularly in the uppermost few thousand feet, and to carry out a more thorough sampling, than in any studies hitherto undertaken.

The flow of heat.—The mean gradient found for the best determinations in the Conglomerate mine is 18.10 $\pm$ 0.23°C./km. To this we add a topographic correction of 3 percent; no appreciable correction is required for the depths (2800 to 5600 feet) under any plausible hypothesis of climatic oscillation. The corrected gradient is thus 18.6 °C. km. The mean value of conductivity for the lavas is 5.01 $\pm$ 0.11 millical cm.sec.deg. The vertical flow of heat by conduction is then 0.93 microcal cm^2.sec.; the statistical standard error of this value is only 2.5 percent, but a more realistic estimate with allowance for inadequacy of sampling might reach perhaps 10 percent.

The mean uncorrected gradient for the Baltic No. 16 drill hole is 16.97 $\pm$ 0.12 °C./km for the interval, 4600 to 6200 feet; with the topographic correction of 3 percent, this becomes 17.5 °C./km. The difference from the Calumet determination of some 6 percent appears to be significant. A difference of mean conductivity of this magnitude between the rocks of this region and those near Calumet seems not unlikely, as the rocks are thought to be more highly altered in the fissured ground near the Keweenaw fault than elsewhere; further study would be necessary to verify this explanation.

The value for the flow of heat near Calumet is in reasonable agreement with measurements of Misener, Thompson, and Uffen (1951; also Leith, 1952) in 11 mines in Ontario and Quebec. Of these results, the best established are those for the mines near Kirkland Lake, Ontario, all close to 1 microcal/cm^2.sec, with estimated probable errors of about 4 percent. These mines are in relatively uniform syenite porphyry, of which 39 samples were used for determinations of conductivity. The workings extend to depths of more than 7000 feet, and below 3000 feet the increase of temperature with depth is well

determined and regular. Most of the other measurements of heat flow in Canada agree within their probable errors, and there seems to be little support as yet for the suggestion of a distinction, based on heat flow, between the Superior and Grenville provinces. Of the three low determinations, that for the McIntyre mine is based on only 3 conductivity samples, that for Malartic Gold Fields on 8 samples and measurements of temperature above 1500 feet, that for the Delnite mine on 6 samples and measurements above 3000 feet. More extensive sampling of the McIntyre mine would be justified, since the temperatures have been well studied down to 5600 feet.

In computing the mean conductivities for various sections, Misener, Thompson, and Uffen (1951) have applied "Chauvenet's criterion of rejection" in a way which seems open to criticism. This criterion is intended to afford an objective basis for discarding erratic values in a sequence of measurements of the same quantity. But in our determinations of conductivity, the experimental error is, we assume, small by comparison with the real differences among samples; a "high" value, for example, does not signify error but the presence of a sample of high conductivity. Our problem is then to determine how much of such material is present and how it affects the flow of heat. If it appears that the exceptional sample is present only in a body of unimportant size, then it may be discarded without further ado; it should not have been included in the sampling. In such questions, there is no substitute for a set of samples which represent, either in their own proportions or by suitable weighting, the material upon which the observed effect depends. It should be recognized, however, that the problem of designing a rational procedure for sampling the many cubic miles of rock whose properties determine the temperature at a point a mile below the surface has not yet been squarely faced, even in principle.

ACKNOWLEDGMENTS

The active cooperation of Dr. Walter S. White of the U. S. Geological Survey was indispensable for this study: Dr. White made available a large collection of samples of rock for the measurement of conductivity, and contributed expert guidance to the geology and petrology of the Calumet region. I am under obligation also to Mr. Harry Donald, Chief Engineer of the Calumet and Hecla Company, for the original determinations of temperature somehow omitted from the paper by Fisher, Ingersoll, and Vivian, and for much other information relative to the location and condition of temperature stations. The unpublished measurements by Alexander Agassiz were obtained from a compilation in the files of Mr. H. C. Spicer, of the U. S. Geological Survey. The samples of serpentine and other minerals were kindly supplied by Professor Clifford Frondel from the Harvard Mineralogical Museum. I am indebted to Mr. Harold Ames for the skillful preparation of all the samples, and to Professor J. B. Thompson for examination of thin sections of the altered lavas.

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[American Journal of Science, Vol. 252, January 1954, Pp. 26-37]

GLACIAL ADVANCES AND SOIL DEVELOPMENT, GRAND MESA, COLORADO

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ABSTRACT. The Grand Mesa, a high plateau in western Colorado, projects far west of the main Rocky Mountain chain. It has been glaciated by three successive ice sheets, leaving two terminal moraines on the top and a younger valley moraine below the rim. The geologic time sequences as interpreted from the location of the moraines are substantiated by differences in the degree of soil development on the three moraines. The first moraine appears to be very old, whereas the last two are relatively young.

INTRODUCTION

Pleistocene events in the Southern Rocky Mountains have received attention in more than 75 papers, beginning with Stevenson in 1875. In 1912 Atwood and Mather called attention to the glacial advances in the San Juan Mountains which lie south of the Grand Mesa. Still earlier, in 1904, Capps and Leffingwell discussed some evidence for a very ancient glacial advance in the Sawatch Range east of the Grand Mesa. In a later paper, Atwood and Mather (1932), dealing extensively with the whole San Juan region, established the Wisconsin, Durango, and Cerro advances as divisions of the Pleistocene and showed the distribution of each on maps. They considered the Cerro to be the oldest and most extensive advance. Atwood and Atwood (1938) summarized information on the Cerro ice sheet in the Rocky Mountains and showed that ice sheets of similar age occur throughout the region. Ray (1940) made an extensive study of glaciation in the Front Ranges and established five subdivisions of the Wisconsin. Wahlstrom (1947) called attention to an extensive ancient ice sheet of early Pleistocene date and described its effect on pre-Pleistocene physiography. There is adequate geologic evidence that the first glacial advance in the Rockies was both ancient and extensive.

The broad phases of the Pleistocene have been worked out for the Rockies, but there remains a wealth of information yet to be obtained by detailed studies. This paper presents the results of a detailed study of glaciation and correlated soil development on the Grand Mesa, Colorado.

The field work was done in 1947 and 1948. All work was done on aerial photos and corrected for distortion when plotted on the controlled base map.

PHYSIOGRAPHY

The Grand Mesa is a plateau in Western Colorado that rises above 10,300 feet. It lies west of the main chain of the Rockies between the converging arms of the Colorado and Gunnison rivers. Its area is 44 square miles. The top is composed of several basalt flows 300 feet or more in total thickness. The basalt, of late Tertiary, rests on Green River shale, but the contact is commonly buried by sliderock. Above the sliderock, the walls are vertical, with few breaks in the rim. Undoubtedly, the basalt field once was wider and extended farther west, but as the Regional drainage system entrenched itself, the weak shales failed under the heavy load, producing Torevalike slides (Reiche, 1937). The mesa is Y-shaped with the stem facing east (fig. 1). Kannah Creek basin lies in the notch of the Y. The north or Lands End branch is about 3 miles wide, while the Indian Point branch is only 1.5

miles wide. The east 2.5 miles of the mesa is a knife-edge ridge produced by block slides on both sides. The westward slope approximates 50 to 60 feet per mile. The slope is steeper at the east end. The elevation at Lands End is about 10,300 feet, whereas the highest east point is near 11,000 feet.

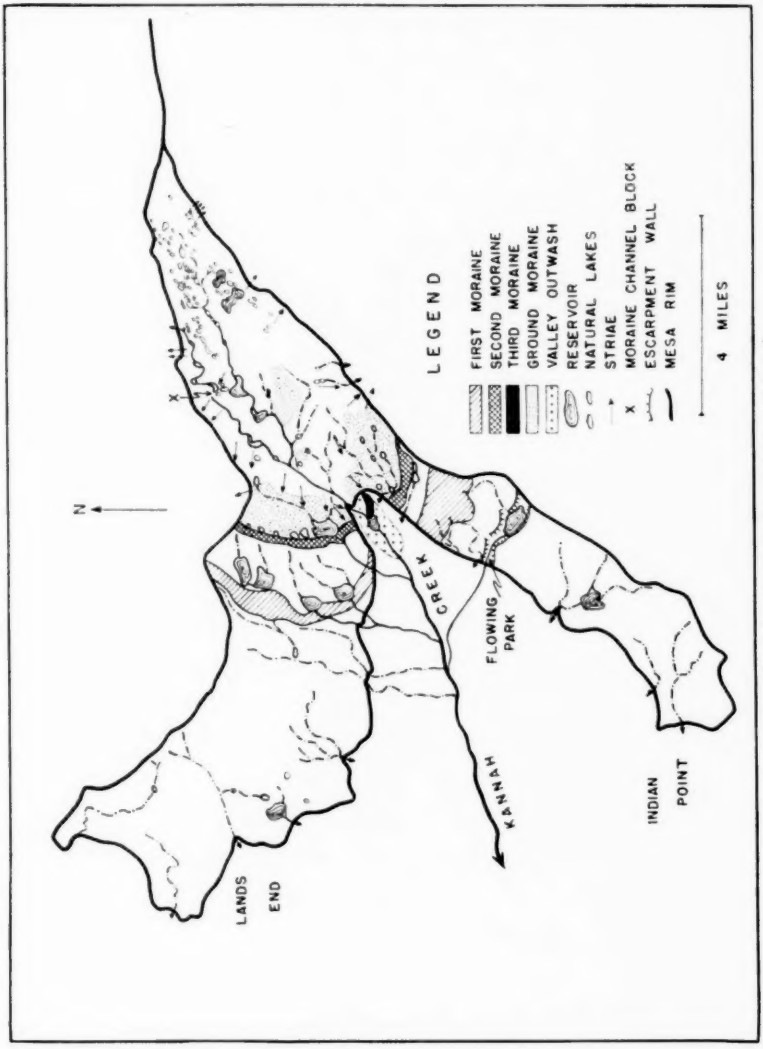


Fig. 1. Moraines on Grand Mesa, Colorado.

Slumping has been very extensive around the entire mesa. The long ridges of successive slides parallel the mesa on all sides. The earlier slide ridges, which are farthest removed from the present rim, are less continuous

and distinct than those near the rim. Similar slides are now in all stages of development on the mesa top from long continuous cracks a few feet across to huge crescent-shaped blocks whose surfaces have dropped 20 to 50 feet below the undisturbed mesa top. The ultimate development will be knife-like ridges. One such ridge now exists at the east end of the mesa and is shown in figure 1.

The western branches of the mesa are open grasslands, but a few miles east of the west end Engelmann spruce grows in patches, and then as solid stands farther east. This dense timber caused some difficulty in studying glaciation details.

The annual precipitation is between 30 and 35 inches. Snow blankets the mesa from September to June. It is said to accumulate thicker above 10,600 feet than below that elevation. A few snowdrifts usually remain in pockets beneath the north rim throughout the summer, but all banks are stagnant and are not considered to be glaciers.

The Grand Mesa is ideally suited to the study of glacial advances because subsequent movements and erosion have not obliterated the deposits on its level top.

GLACIATION

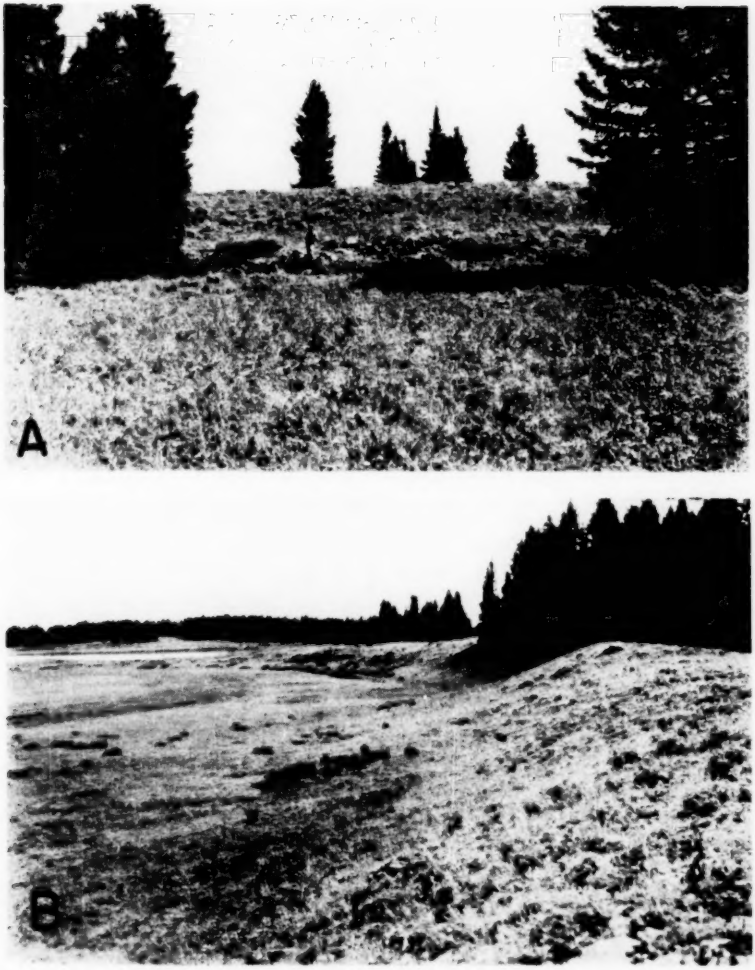
Moraines.—Henderson (1923) reported on the glaciation of the Grand Mesa. He mentioned that about half of the top had been glaciated, but his report was primarily concerned with glaciation beneath the rim on the north and east (?) sides. The moraines referred to appear to be somewhat confused with Toreva-block slides.

The mesa top was subjected to at least three glacial advances, two of which left terminal moraines on top. The third left a valley terminal in Kannah Creek.

The first or oldest terminal moraine, at about 10,500 feet elevation, is best developed on the Lands End side of the mesa (fig. 1). In places this ridge is about 30 feet high and 1/4 mile wide but elsewhere, and especially on the Indian Point side, the front and back faces are less distinct. Its general lack of sharp form could be attributed to a single advance followed by rapid retreat, to erosion modification due to its antiquity, or to a high proportion of silt and clay compared to boulders in the till which permitted the slopes to assume low angles. The terrain over which this ice sheet moved was presumedly deeply weathered, and therefore the proportion of silt and clay would be higher than in later till deposited by ice sheets moving over a surface previously scraped free of weathered material. This moraine contains fewer large boulders and more fine materials than the second moraine. There are no kettles.

Meltwaters from the Indian Point side drained into Kannah Creek basin through Flowing Park, a shallow basin which cuts directly across the mesa top. On the Lands End side meltwaters developed four drainageways, all emptying into Kannah Creek. Three of these later dissected the moraine and drained meltwaters from the second advance.

PLATE 1



The second terminal moraine on Grand Mesa.

A. Distal view showing abrupt slope and common height.

B. Proximal view showing abrupt face with marshes and lake against moraine.

The second and most prominent moraine lies approximately 1.25 miles east of the first and 75 feet higher in elevation. It averages about 0.25 mile wide. Both its distal (pl. 1, A) and proximal (pl. 1, B) sides are distinct and sharp. The sharpness of the proximal side can be attributed to ice banks and to direct shove. On both branches of the mesa, this moraine is thickest or deepest adjacent to Kannah Creek where it is about 30 feet high. It decreases to about 5 feet high at the north and south rims. As in the case with the first, the second moraine terminates abruptly on both sides of the present rims of Kannah Creek. The surface of the second moraine is extremely irregular. Kettles are 5 to 10 feet deep. The till is very stony. The ice sheet may have stood for a considerable time at this point adding material, or it may have been carrying more boulders and less silt and clay than the first moraine.

Between the two terminal moraines, the ground moraine is obscured. Outwash from the second undoubtedly filled some of the original kettles and depressions behind the first. The first moraine on the Indian Point side (fig. 1) includes some outwash material. No attempt was made to separate outwash from till on this side since the outwash is not clearly distinguishable by relief features. On the Lands End side, the outwash plain in front of the first moraine is indistinct. Deep areas of ground moraine are shown in figure 1. Eastward where the mesa top was scraped and gouged deeply, only small areas of ground moraine occur and are not shown in figure 1. Judging from an area recently exposed by fire, the percentage of exposed bedrock is high in this east end.

The Pleistocene was brought to a close by localized movements down valleys on the mesa as indicated by striae and gouges in the rocks on these valley sides. Many of these valleys have been carved from solid bedrock and the markings are clearly preserved. The mesa portion of the Kannah Creek channel enters the basin of Kannah Creek accordantly about 1/3 mile east of the second moraine and at this point the cross section is not the idealized broad U-shape of some glaciated valleys. Instead, it has two distinct sets of terraces. The first pair lie 10 to 15 feet below the top of the mesa and are roughly 1/3 mile apart. The west member, in its lower extension, is a bare rock bench terminating in a vertical wall a few hundred feet east of the second moraine on the Lands End side. This bare rock is marked by glacial striae. The second set of terraces is about 1/5 mile apart and about 20 feet below the first set. The present channel of the creek lies between and well below the terraces. This arrangement extends upstream for some distance and then fades out as the creek bed rises, becomes more shallow, and then becomes peat-filled (pl. 3, B).

The presence of a prominent, curved, 40-foot-high terminal moraine—the third moraine—1/2 mile below the entrance of Kannah Creek into the basin, represents the end of an advance in the basin (fig. 1). The valley center consists of outwash deposits reworked by Kannah Creek.

Other valleys carved in rock are scattered over the mesa top. They are not as deep as the lower extension of Kannah Creek and have more of the expected U-shape. Some have falls in their lower reaches, apparently due to ice plucking. Some of these valleys pass directly over the rim without the forma-

tion of prominent rock cones or plunge pools, while others have built up a cone against the walls.

Glacial markings.—Striae and gouges are well developed (pl. 2, A). Because of weathering, many of the finer markings have been obliterated, and others can only be seen by careful observation. The larger markings are clearly visible and readily show the course of the ice movements. Some prominent striae were plotted on aerial photographs and their directions transferred to figure 1. They indicate three general movements. The first set of striae are perpendicular to terminal moraines. These marks occur in the rock on both sides of Kannah Creek, indicating the ice overrode the creek if the creek existed at the time the ice sheets moved down the mesa. The second and third groups clearly show that ice moved outward and over both the north and south rims at many places. The marks frequently and abruptly end at the very outer edge above the rim, although there is no catchment basin or evidence of any kind to indicate that ice once went over the rim in its present location.

Roches moutonnées (pl. 2, B) are present in many places, but are difficult to recognize on the far east end because of the dense timber. They are most conspicuous just east of the ground moraine. None occur between the two terminals.

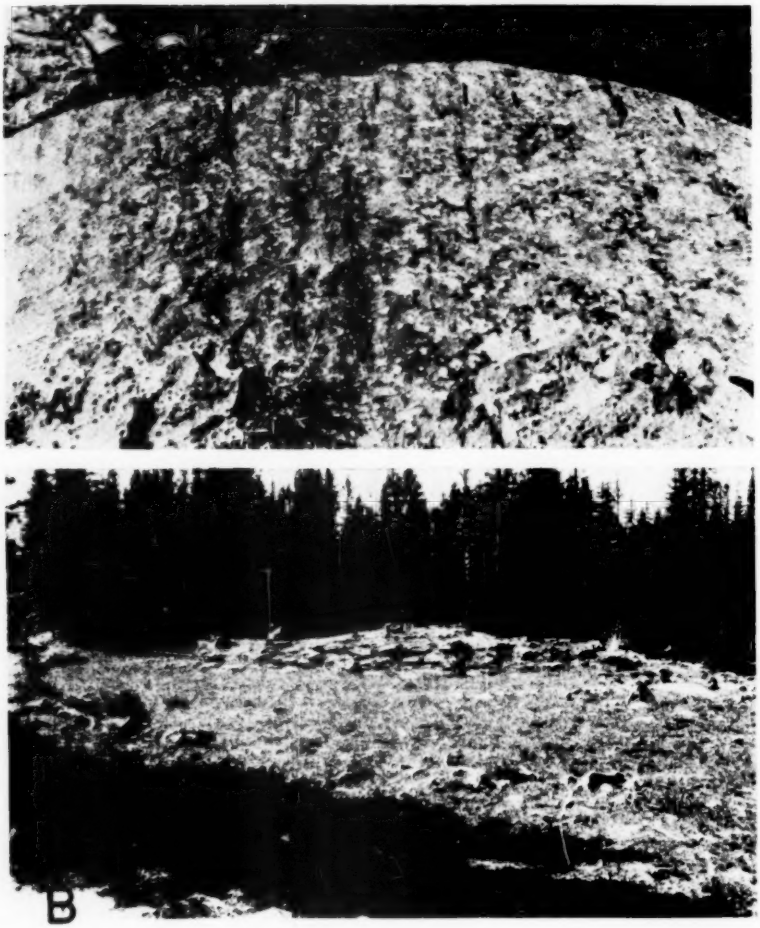
Lakes and kettles.—The Grand Mesa region is noted for its many lakes, said to be about 400 in number. Many of the larger ones lie beneath the rims in the elongated Toreva-like block basins paralleling the rims. The mesa top is spotted with dozens of large and small lakes. Some have been enlarged for irrigation reservoirs and some drainageways have been dammed for reservoirs. The natural lakes occupy kettles and basins gouged out of the rock. They range from mere peat-filled depressions to bodies of water several acres in size (pl. 3, A). Many are round; some are oblong; and others are irregular. There is no consistent orientation of the oblong lakes. The lakes are concentrated on the east end where the slope is steeper and presumedly the glacier was cutting most.

Many kettles contain no water. Kettles are absent in the first moraine, perhaps having been obliterated by erosion or outwash. Ancient kettle lakes behind this moraine may have been drained later by streams that dissected the moraine. Kettles are numerous in the second moraine.

Streams.—Many of the streams that drain the mesa empty into Kannah Creek basin. Drainageways are much less well developed on the western nonglaciated portion. On the Indian Point side, the meltwaters of the first moraine emptied into the basin in Flowing Park and two small streams east of Flowing Park, but drainage of the second moraine developed parallel to its back side and now drops over the rim into Kannah Creek without cutting through the moraine. The streams behind the second moraine on the Indian Point side were diverted from their original courses by both the terminal moraine and by local ground moraines.

On the Lands End side, the first moraine is cut by three streams, and two others drain its face along the north side. All are more deeply entrenched toward the mesa rim than are those streams farther west. The second moraine is cut by only one stream, although several springs occur along its face.

PLATE 2



A. Glacial markings as they commonly occur in the basalt rock. Pencils indicate the larger gouges.

B. One of the many well-developed roches moutonnées east of the second moraine. Ice moved from right to left.

East of the moraines, Kannah Creek dominates the drainage (pl. 3, B). It is deeply cut in the basalt. At point X in figure 1, its channel was blocked with till from an ice sheet moving northwest and waters now drain east and west from this obstruction. The odd zigzag shape of the eastern part of this stream is difficult to explain. The channel of this section is 30 to 50 feet deep and, since it has carried a glacier, one could have expected the sharp angles at the turns to have been rounded off. There is some evidence, as shown by striae, that the ice did not make these turns, but rode out and over the rim from the points of the north-facing angles.

SOIL DEVELOPMENT

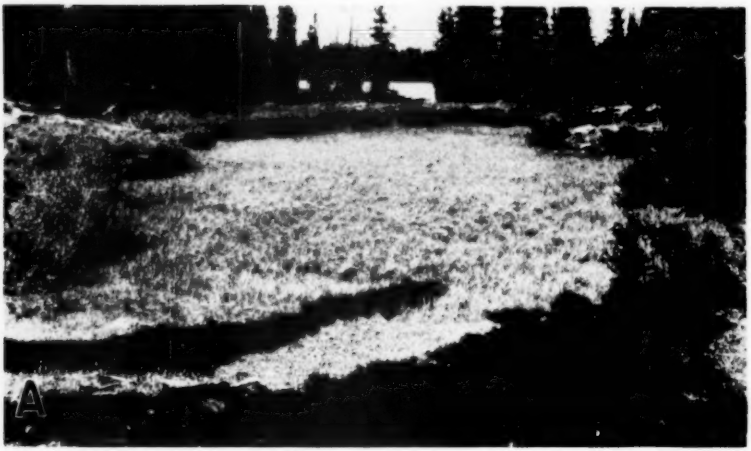
The Grand Mesa is a unique place to study soil development. The genetic processes are not obscured by many unknown variations that cannot be reasonably evaluated. First, the parent rock—basalt—is lithologically uniform. The topography has low relief, best characterized as gently undulating. The drainage pattern is very open and channel entrenchment is at a minimum. The evidence indicates that surface erosion over geologic time has been limited. Since the mesa is of late Tertiary age, the time factor is doubtless adequate for the pedologic processes to have fully expressed themselves. It is not presumed that either the vegetation or climate have remained the same throughout this period, but certainly, because of its limited size, all changes have affected the entire mesa equally.

Since essential uniformity is inferred for the soil-forming factors, the mesa top should have one type of profile, provided no major event occurred to interrupt the processes. But major interruptions did occur, as shown in the preceding discussion of glacial advances. Because of the advances, important differences are now reflected in the soil profiles. The profile characteristics throw some light on the relative dates of the different advances.

The soils have been studied in detail. West of the first moraine on the nonglaciated portion of the mesa, the profile has a dark surface soil and a clearly marked reddish or yellow-brown subsoil. This subsoil has a blocky and prismatic structure which changes to massive tough clay with a redder color below 30 inches. This layer grades down into the bedrock and in roadbanks it can be seen filling crevices as deep as 10 to 15 feet. The pH values are about 6.5 for both surface and upper subsoil. This distinctive or "regional" profile can be considered the summation product of all the forces of the environment, including the many variations that have occurred since the Tertiary. There is no known method to evaluate or to segregate profile modifications attributable to major environmental changes during this period, even including the relatively late influence of the present environment. It is not considered to be a senile or overmature soil, although the criteria for determining this are poor indeed in the present stage of pedologic science.

The advance of the first moraine destroyed the regional soil profile over the area of its advance. Thus, the soil-forming processes would necessarily have to start again with till in which there was no semblance of a profile. Soil development took place in this old till and the profile that is now present

PLATE 3



A. Gouge and kettle lakes on east end of Mesa. The marsh in foreground is a typical example of the most common type.

B. Kannah Creek on the Mesa top. The bed is peat-filled. The sides and local topography are strongly modified by ice movement.

in the first moraine is nearly identical with the regional profile on the west nonglaciated part. This has three possible explanations: (1) soils develop rapidly in basaltic materials; (2) the environment, particularly the precipitation and temperature, was very favorable to soil development following this advance; (3) this early advance is very ancient, permitting an adequate extension of the time factor.

The profile in the second moraine, in the ground moraine behind it, in the third or valley moraine, and in all the eastern part of the mesa, has no clay subsoil. The dense clay layer has not developed. The comparison of mechanical composition between these two profiles is shown in table 1. This recent profile is outstanding in its lack of subsoil morphology and clay content. As above, the possible explanations include a climate unfavorable to soil development or the lack of adequate time.

TABLE 1

Comparison of the Mechanical Composition of the Regional and the Recent Profiles

Soil Profile	Horizons and Depth		Gravel and Rock > 2 mm	Sand 2.0–0.05 mm	Silt* 0.05–0.002 mm	Clay < 0.002 mm
	Inches		*Percent*	*Percent*	*Percent*	*Percent*
Regional profile	A_1	0–18	0.9**	35	61	4
	A_3	18–28	0.8	38	54	8
	B_1	28–34	0.8	38	45	17
	B_2	34+	40.1	35	35	30
Recent profile	A	0–18	7.3	36	56	8
	B	18–34	43.9	38	54	8
	C	34+	43.5	54	40	6

* Analysis by hydrometer.

** Gravel calculated as percentage of entire sample and the sand, silt, and clay as percentage of the < 2 mm fraction.

DISCUSSION

Because of the flat terrain of the Grand Mesa, the deposits left by glacial advances have been well preserved. The first or oldest moraine is doubtless of pre-Wisconsin age because it is the most westerly deposit, its topography is subdued, and the soil profile is far advanced in subsoil development.

Atwood and Mather (1932) referred to the pre-Wisconsin advance in the San Juans as Cerro till, while a second and younger advance was referred to as Durango till. The age of the second moraine on the Grand Mesa, from its position, might be considered of Durango age. However, the relief of this moraine is sharp and unsubdued and the soil profile shows no evidence of subsoil development. Therefore, this moraine is considered to be relatively young.

The evidence is clear that both the first and second moraines were continuous and that Kannah Creek basin had not at those times extended itself as far east as its present location. The distance of the headward erosion of Kannah Creek basin into the basalt cap is about the same between the first

and second moraine as between the second moraine and the present head of the basin. This could be construed as evidence that the second moraine is of intermediate age between the first advance and the third advance which left the valley moraine. However, distances of retreat may not be as important as the quantities of water available for headward erosion at different times in the Pleistocene. From the lack of soil development in the second moraine, it is concluded that this moraine is of Wisconsin age and not of Durango age unless the latter can be considered as early Wisconsin.

The third or valley moraine is doubtless of Wisconsin age. It was deposited by ice moving down from the mesa top through Kannah Creek channel. Striae on the valley walls clearly show this movement. Subsoil development is lacking in all of these deposits, including the ground moraine and the glaciated portions of the mesa behind the moraines. This suggests that the time difference between the second and third advances has not been great enough to result in differences in profile developmnt.

Although Ray (1940) found evidence for five substages of the Wisconsin in the Front Ranges, this study will support only two advances. All these deposits occur on flat terrain where a moraine would not be destroyed except by being overridden.

It is evident that the area of the mesa top has been reduced since the Pleistocene. The headward development of the Kannah Creek basin has removed large sections of both the first and second moraines. Ice sheets passed out over the mesa rims above the moraines as demonstrated by striae. In many places the striae occur at the very edge of the present rim and yet there are no cirques, plunge pools, or debris cones to mark the flow of the ice over the rim. This can only mean that the rims have retreated since the ice passed over.

The entire mesa is flanked at its base by a series of long, crescent-shaped rock ridges paralleling the face of the rims. These occur in a series outward from the rim with each succeeding and lower ridge losing more sharpness of outline and continuity from end to end of the crescent until all individual identity is lost in rock fields or becomes jumbled masses of rock. It is clear that these ridges are remnants of huge landslides resulting from failure of the shales beneath the basalt cap when an abundance of water lowered the competency of the shales. In places, one end of the last slide at the base of the rim can be traced upward to the top of the mesa. It has been mentioned that future slides in all degrees of development from wide continuous cracks to masses whose surfaces have already dropped well below the top occur along the mesa top today. The ultimate result of this wasting of the land mass by slides will be long knife-edge ridges as is so well illustrated by the ridge on the east end of the present mesa (fig. 1).

From the above, it is clear that all evidence of the ice sheets that passed over these rims have been enveloped in this great mass movement of land slides until traces of till or glacial land forms are no longer identifiable as such.

It is hazardous to estimate the distance of retreat of the mesa top since the beginning of the Pleistocene. But it may be pointed out that such a dis-

tance might be approximately half the present distance between the moraines where Kannah Creek has severed them at the head of the Kannah Creek basin as shown in figure 1.

The relation of soil development to geologic events is striking in this locality. The contrast is great between the strongly developed subsoils in the first moraine and the lack of subsoils in later deposits. The contrast is less marked and subsoil development is nearly identical between the first moraine and the nonglaciated area to the west. The period of time since deposition of the first moraine must have been great enough to permit this soil development. This supports the antiquity suggested for the pre-Wisconsin deposits.

The author is familiar with a number of the localities in Colorado where Cerro till has been described by Atwood and Atwood (1938). In each locality these deposits are distinguished by soils having well-developed subsoils very similar to those found on the Grand Mesa. Likewise, Hunt and Sokoloff (1949) have called attention to soil profiles that distinguish pre-Wisconsin deposits in the Rocky Mountain Region.

It appears to the author that the sciences of pedology and geology have much in common and that great forward strides can be made by complementary studies.

ACKNOWLEDGMENT

The author gratefully acknowledges suggestions from Professor R. F. Flint which have been helpful in preparation of this paper.

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ROCKY MOUNTAIN FOREST AND RANGE EXPERIMENT STATION
FOREST SERVICE
U. S. DEPARTMENT OF AGRICULTURE
FORT COLLINS, COLORADO

[American Journal of Science, Vol. 252, January 1954, Pp. 38-54]

DOES THE COLD OF WINTER AFFECT THE THERMAL INTENSITY OF THE HOT SPRINGS IN YELLOWSTONE PARK*

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ABSTRACT. Because of their location Yellowstone's hot springs are subject to extreme seasonal climatic changes. This has led to a theorizing that hydrothermal functioning would be less intense in winter than in summer. This paper presents quantitative data taken from summer and winter observations on temperature and function of many of the hot springs in the Upper Geyser Basin, so that comparisons might be made in determining if there are seasonal variations in the intensity of the thermal energy. The temperature determinations cover most of the important springs located in the Upper Basin. The statistical data on hot spring activity are largely confined to the larger eruptive units, whose functions follow a pattern of activity that would lend itself to a comparative study. Some of the factors known to influence hot spring behavior, which might lead to a misinterpretation of data, are briefly considered.

INTRODUCTION

Observations of Yellowstone Park's hot springs since their discovery in 1870 have been confined almost wholly to the summer season. With the exception of Old Faithful Geyser few or no studies of a scientific character have been made of geyser activity during the winter months. These hot springs are located in one of the coldest sections of the United States, a fact which has given rise to a frequent query as to whether or not the cold of winter might measurably retard the functioning of some of the eruptive springs. That winter might prolong the length of the eruption intervals for at least some geysers was made plausible by the behavior of the Daisy Geyser on windy days. During the history of the Daisy all who have observed its eruptive pattern over a few weeks' period have become aware that the length of the average eruption interval is increased as a result of the wind; the stronger the wind the greater the delay. It was explained that convection carried the surface water, chilled by the wind, into Daisy's well, necessitating a longer period than would otherwise be the case to heat the water to the critical temperature of an eruption (Lystrup, 1934, p. 26). The same explanation has been given as one of the reasons for a lack of activity of the Great Geyser in Iceland when winds are blowing over its surface.

If the eruption interval of a geyser can be measurably lengthened by increased surface cooling in summer, it would seem that subzero conditions, which are so common in the geyser basins of Yellowstone Park from November until April, would not only greatly aggravate this condition of delayed activity but could, theoretically at least, produce quiescence in the Daisy and all geysers of its type. Prior to the record which was kept of the Daisy during the 1951-1952 winter there is no evidence that statistical data on its winter activity suitable for comparative study were ever obtained. One object in carrying out the 1951-1952 winter observations was to supply these data so that intelligent deductions would be possible in determining if increased

* Report on winter observations of hot spring activity in the Upper Geyser Basin for the winter of 1951-1952, and for November and December 1952. Published with the permission of Conrad L. Wirth, Director, National Park Service, Washington, D. C.

surface cooling resulting from the winds were a true explanation for the Daisy's less frequent performance on windy days. Should surface cooling affect its eruptive pattern, it would be logical to assume that other geysers which have inverted cones would also be responsive to these same external conditions and that a winter study would determine if there were seasonal variations in their behavior. It was further believed that temperature studies, in connection with detailed records of geyser activity, would furnish sufficient data for valid conclusions.

WINTER OBSERVING DONE PRIOR TO 1951

From 1925, and for the succeeding seven seasons, Dr. E. T. Allen and Dr. Arthur L. Day, as representatives of the Geophysical Laboratory, made intensive studies of Yellowstone's hot springs. It seemed desirable for them to supplement their summer studies by winter observations. To accomplish this they sought the cooperation of the National Park Service, which resulted in the stationing of a park ranger in the Upper Basin during the winter months. In commenting on the need for winter studies they state: "Up to the time of the present investigation (1925-), scientific observation in the Yellowstone Park has been entirely confined to summer and early autumn, but as no satisfactory basis for a hot-spring theory appeared possible without some knowledge of seasonal influences on temperatures and geyser activity, some arrangement for winter observation seemed imperative. Accordingly a definite program of observation was undertaken at the Upper Basin by permanent Park Rangers, in connection with their regular duties in protecting wild life" (1935, Preface, IV).

This program of observation was carried on for five successive seasons, from 1925 to 1930. Allen and Day report that the park rangers who were stationed in the Upper Basin during this period were all agreed that so far as their observations went, the season of the year had no perceptible effect on spring discharge or geyser activity.[1] It should be stated that these conclusions were very largely based on subjective premises. There is no evidence that a continuous daily record of any geyser's activity was kept by the rangers, or that summer records of sufficiently detailed character were available, which records would be imperative if comparative data of a scientific character were to be made. Further, during each succeeding winter a new ranger was assigned to the task. Park Ranger R. M. Baker spent the longest period, from October to May 1928, "and again the next season till February 1929." These short periods of observation could hardly suffice to acquaint the rangers with the idiosyncrasies in behavior which a number of the geysers show. Without accurate summer data or detailed personal knowledge of a hot spring's behavior, and in the absence of a quantitative winter record, observations such as were made during the 1925 to 1930 winters are little more than generalizations.

[1] Reports now on file in the Superintendent's Office made by the rangers do not lend support to the conclusion that geyser activity and discharge are independent of the season of the year. Allen and Day's conclusions of no seasonal changes in thermal intensity would seem to have been wisely based on comparative temperature studies.

Comments of Allen and Day on the results of the 1925-1930 winter observations express satisfaction with the results obtained by the park rangers. The authors draw rather definite conclusions in interpreting hot spring activity from the winter observations, yet their summarization clearly reveals that these observations on geyser behavior were wholly subjective, and hence of little scientific value in formulating any geyser theory. "As to the results, all the observers were agreed that, so far as they could decide from qualitative observations, neither spring discharge nor geyser activity was perceptibly affected by weather or the season. As regards *pronounced* seasonal variations, it is believed that the judgment of these men, carefully selected for this work, may be safely trusted. Smaller systematic differences, to be sure, might be overlooked by an observer who depended entirely on estimates" (Allen and Day, 1935, p. 246). It is these smaller "systematic differences" that are of paramount importance if accurate determinations in seasonal variations are to be made. "Qualitative observations" of geyser activity have little scientific value when it is measurable variations in comparative data that are sought.

During the winter of 1949-1950 District Ranger Ruben Hart was stationed in the Upper Basin. Being of a scientific bent, Mr. Hart, with the cooperation of the Naturalist Division, by using an "automatic geyser recorder," secured an unbroken record of 2605 eruptions of Old Faithful Geyser. The record covered a period from "November 30, 1949 and continued through April 30, 1950." From this record Hart made some interesting statistical observations (1950, p. 37). Old Faithful was the only geyser he studied. Apart from monthly temperature records of many of the hot springs made by the park rangers stationed in the Upper Basin from 1925-1930, it can safely be said that Hart's report was the first scientific contribution to a winter study of any geyser in Yellowstone Park.

TYPES OF GEYSERS STUDIED

Two classes of geysers are recognized in Yellowstone Park; they are referred to as *cone* and *fountain* in type. In general the cone type geyser, Old Faithful being representative, is nozzled down to a small aperture, so that only a small amount of its water is exposed to the atmosphere. The fountain geyser is an inverted cone with a large surface exposure of water resting above the constricted portion of its orifice. The cone geyser is in a large measure insulated against wind and cold and theoretically should be largely immune to influences from these factors. If measureable influences from cold and wind are to be detected, those geysers with a large surface exposure to the atmosphere, such as the Grand, Oblong, Daisy, etc., should be the ones where such influences would most likely be manifest; hence it was to this latter class of geysers that special effort was directed in the winter observations begun November 1, 1951. The cone type geysers were not overlooked; the eruptive patterns of many were checked with the same care as were those of the fountain geysers.

One of the seeming disadvantages of a winter project of observation such as was started in 1951 is that to make the data conclusive several seasons of continuous observation — winter and summer — are required. With but

one or two winters of observation it is not possible to eliminate completely those influences due to a shifting of the thermal energy from one spring to another. Exchange of function is the biggest single factor in producing the great irregularity of many geysers (Marler, 1951, p. 329). Without an appreciable time range in data, the influences resulting from a shift of the thermal energy might become confused with the season and interpreted as weather influences.

Cold is more intense in the geyser basins some winters than others. It is possible that prolonged and severe cold could result in retardation of the function of some geysers, a slowing up which would not be measurable with the average winter temperature conditions in the basins. So far as temperatures were concerned, the 1951-1952 winter was unusually benign in Yellowstone Park. No periods of extreme cold, such as December 1924 or January 1936, were even closely approached.

COMPARATIVE DATA, SUMMER AND WINTER, ON SOME OF THE GEYSERS

In presenting the data collected on the geysers that were under observation, space in this paper will not permit the showing of more than average eruption intervals for the periods covered by the comparisons. The detailed statistics on daily temperature, weather, and eruption times are on file in the Yellowstone Park Library. The comparative eruption data, with explanation of factors other than weather which are known to influence the eruption times of many of the geysers, should, however, give a substantial basis for drawing conclusions on whether winter influences geyser activity.

Daisy Geyser.—The Daisy lies about one mile northwest of Old Faithful. Its high degree of regularity and frequency have contributed to making it one of the most popular and better known of the geysers. Over the greater portion of its history only Old Faithful among the major geysers has exceeded it in the frequency of its eruptive periods. Unlike Old Faithful, the Daisy is known to have subterranean connections with other thermal units.

On July 10, 1951 the Splendid Geyser, which has underground connections with the Daisy, became rejuvenated after over 50 years of dormancy. This resulted in very erratic functioning of the Daisy for the remainder of the season. It was feared that Splendid's activity would make it impossible to get enough of Daisy's winter function for comparative data. Fortunately, from November 9 until December 20 it played without any apparent influence from connecting units. During this period 140 eruption intervals were checked. The average interval was 108.2 minutes. From May 1, 1951 until July 10 when the Splendid became active, the average eruption interval was 105 minutes; from July until October 1, during the period of Daisy's erratic function, it was 140 minutes.

From December 20, 1951 until January 1953 the Daisy was inactive the greater part of the time due to the activity of either the Splendid Geyser or Brilliant Pool. During the November-December 1951 period of activity the longest interval checked on the Daisy was on November 29. This interval was 130 minutes; the wind was 20 miles per hour from the southwest; the tem-

perature 30° F.[2] The shortest interval, 85 minutes, was on November 24, at which time the air was calm and the temperature 30°. The highest temperature recorded at the time of an eruption was on November 30 and was 41°; the coldest was on February 24 and was −28°. Despite a loss of 69° in temperature, the eruption on February 24 was on the same length interval as the one on November 30—105 minutes. At the time of both eruptions there was a slight breeze from the north.

Giant Geyser.—During the 1951 season the Giant was unusually active. From May 1, when the observations began, and for the remainder of the year, the Giant played with greater regularity and frequency than during any previous observation period for which there is a record.

In the middle of January 1952 the Giant suddenly began playing on even shorter intervals than had characterized it during the 1951 season. From January 17, 1952, and for the remainder of the year, it played with unprecedented frequency and regularity. The average eruption interval during this period was slightly over 2 days—57½ hours. The extreme variations from this average were 88 and 34 hours. Eighty percent of the eruptions came within 10 hours of the average. The Giant has the distinction of being the first geyser in Yellowstone Park with a complete yearly determination of its activity. During 1952 it erupted 137 times.

The Giant is definitely a cone type geyser; it is connected underground with over 50 other known units. Some of these units, such as the Purple Pools, have large surface exposures. In spite of the great surface area of water which is known to belong to its plumbing, its regularity was undisturbed, so far as could be determined, by winter conditions. It is noteworthy that the shift[3] from a plus 5- to a plus 2-day average interval occurred in mid-January. This would scarcely suggest a delayed function due to the cold of winter.

Grand Geyser.—Although the functional characteristics of the Grand are very distinct from those of the Daisy, their surface conditions bear marked similarities. Both have large surface areas exposed to the atmosphere; both are in a state of overflow for about the same period prior to an eruption. In spite of this surface relationship there has never been any suspicion that winds were a factor in delaying the Grand's activity. If surface cooling delays geyser activity it would seem that all open crater geysers like the Grand and the Daisy would be influenced in some measure by external temperatures.

Not only do the Grand's eruptive periods seem to be uninfluenced by any degree of wind but they also show an equal independence in function with respect to the season. Its eruptions appear to come with as great a frequency when subzero conditions prevail as during maximum summer temperatures. The eruption data on the Grand for the 1951 summer when compared with the data collected during the following winter should bring out this fact very forcefully. From May 25, until September 15, 1951, 136 eruption intervals were determined. The average eruption interval was 20 hours. The minimum interval during this period was 11 hours, the maximum 38 hours. From November 9, 1951 until April 28, 1952, 208 eruption intervals were deter-

[2] All weather temperatures given in this paper are Fahrenheit readings.

[3] For cause of shift see Yellowstone Nature Notes, Vol. 26, No. 6, page 67.

mined. The average interval was 14¾ hours. The maximum interval during this period was 22 hours, the minimum 8½ hours. Some of the shortest intervals during the winter were at the times of subzero weather. The coldest temperature recorded at the time of an eruption was on February 24; it was −32°, and the interval was 13 hours, this being nearly 2 hours under the winter average.

Starting on November 22, 1952, an arctic mass of air moved over the geyser basin, and for 10 days the temperature ranged well below zero. The average minimum temperature during these 10 days was 17.4° below zero. During this period 12 eruption intervals of the Grand were determined; the average time was 13.3 hours. For the preceding 10 days, from the 12th to the 21st inclusive, 10 eruption intervals were determined, with an average interval of 15.5 hours. The average minimum temperature during this period was 14.6° above zero.

Castle Geyser.—All observations to date indicate that the Castle is a self-contained unit, although its functional history suggests influences from subterranean connections with another spring or springs. For long periods it will show a degree of regularity on more than one eruption schedule; at other times it becomes very irregular, infrequent, and unpredictable. This behavior is characteristic of all geysers known to have underground connections with other units.

During all of the 1951-1952 winter the Castle showed a high degree of regularity and frequency. From November 5 until April 26, 125 eruptions were observed; 89 eruption intervals were determined. For from 5 to 7 days the intervals would vary from about 12 to 13½ hours, then one interval of about 20 hours would intervene to be followed by the shorter intervals for the period indicated. The 13.1-hour average for the winter is shorter than any summer average for which there is a record. It continued playing on the 13.1-hour average until July 10, 1952, following which date, and for the remainder of the year, the average eruption interval was 14¾ hours.

Oblong Geyser.—During the 1951-1952 winter no detailed data were kept on the Oblong. Frequent observations indicated a pattern of play comparable to its summer performance. From November 1, 1952 until January 1, 1953 an effort was made to record as many eruption intervals as possible. During this period 12 such intervals were determined. The Oblong is a fountain type geyser with a large area exposed to the atmosphere. This, however, was no deterrent to the frequency of its winter function. Of the eruption intervals checked, the longest was 4 hours; the shortest 2 hours and 55 minutes; the average, 3.4 hours. For 11 eruption intervals checked from June through August 1952 the average interval was 3.7 hours.

The Oblong, like several other springs in the Upper Basin, shows effects from spring drainage; the water in its bowl becomes about the same hue as the spring runoff in the Firehole River. The Oblong's crater rests on the bank of the Firehole, the crater rising about 10 feet above the river. Whether at this season of the year the humus-colored water enters the Oblong's water supply directly from the river or represents surface drainage has not been determined. Whatever the source, it acts as no apparent deterrent to the

Oblong's play. In May 1951 while it was clouded with brownish water, an eruption interval of 2 hours and 58 minutes was determined. If the melting snow water does not affect the function of a geyser like the Oblong, it is scarcely plausible that atmospheric temperatures would have any effect in slowing eruptions.

Riverside Geyser.—The Riverside is a cone type geyser, and is perhaps the most regular of the large geysers in Yellowstone Park. Its small surface exposure, like Old Faithful's, should protect it from the atmosphere. As has been pointed out, geysers like Old Faithful and Riverside might be expected to play as frequently and regularly in winter as in summer if atmospheric temperatures were the only factors involved.

During the 1951 season, from May 1 to October 1, the Riverside played 495 times. The average eruption interval was 7 hours and 25 minutes. From November 9, 1951 until April 27, 1952, 572 eruptions took place: the average time was 7 hours and 3 minutes. The shortened winter average persisted during the 1952 season, being an even 7 hours. During most of the years that the author has had the Riverside under observation its seasonal average has been near 8 hours.

Beehive Geyser.—The intervals on the Beehive were carefully checked during the winter observations. Its eruptions do not come with sufficient frequency and regularity to serve as a possible index to atmospheric influences, but its performance is spectacular and all of the eruptions could easily be determined.

References to the Beehive during the first two decades of the Park's history indicate that there were some seasons when it played one or more times each day. During the greater part of the present century it has been relatively inactive, and for 25 years, starting about 1920, many seasons went by with no observed activity. Starting with the 1947 season there was a general increase in geyser activity in the Firehole Basins, in which the Beehive shared. Since the 1947 resurgence it has played on frequent occasions. Starting with 1947 the number of eruptions checked per season were: 1947, 26; 1948, 13; 1949, 28; 1950, 23; 1951, 23; and 1952, 31.

Early in November, 1951 it became apparent that the Beehive was playing with unusual frequency. From November 8, when the recordings began, until April 16, 1952, the Beehive played 113 times. The numbers of eruptions per month were as follows: November, 18; December, 24; January, 22; February, 20; March, 18; April, 11. From May 12 to September 3, 1952, 31 eruptions were recorded. The eruption on September 3 occurred soon after the beginning of an eruption of the Giantess. This eruption of the Giantess completely upset the normal activity of most of the springs on Geyser Hill, including the Beehive. From the September 3 eruption it was dormant until November, when its winter rejuvenation started, with 8 eruptions; during December there were 21, and 3 more occurred up to January 4 when the observations ended.

Old Faithful Geyser.—Since a careful and scientific check had been made of Old Faithful during the winter of 1949-1950, the 1951-1952 winter observations were directed toward other geysers in the basin. Some of the

random observations made over the years were indicative that Old Faithful's pattern of activity was essentially the same in winter as in summer. Ranger Hart's automatic recordings bore out these generalizations. During the 1949 season 1540 eruption intervals were determined: 35 minutes was the minimum interval, 88 minutes the maximum; the average eruption interval for the summer, 63.2 minutes. During the following winter 2505 eruption intervals were mechanically determined. The winter record showed a minimum interval of 38 minutes, a maximum of 88, with an average of 63.8. The winter record when compared with the summer one shows an increase of .6 of a minute; the maximum interval, however, for both summer and winter is the same. That the .6 of a minute in the average gain of winter over summer was not due to seasonal influences is indicated by the record of the following summer which shows an average of 63.8 minutes for 1250 eruption intervals, the same as the previous winter average.

The winter observations begun in November 1952 also reveal Old Faithful's independence of outside temperature. The summer record for that season shows that 1289 eruption intervals were determined; 35 minutes was the minimum, 85 minutes the maximum; the average, 64.09. From November through December 1952, 619 eruption intervals were recorded: 39 minutes was the minimum, 77 minutes the maximum; the average, 63.6—.49 minutes less than the summer average. From November 22 through December 1 the temperature in the Upper Basin ranged between 5° and 32° below zero. The time of the average eruption interval during these 10 days of subzero weather was 63.59 minutes.

WHAT TEMPERATURE STUDIES SHOW

During the 1925-1930 winters, Rangers Phillips, Baker, Hanks, and LaNoue made monthly temperature determinations of a number of hot springs in the Upper Basin. Allen and Day's studies of these temperatures indicated that "in winter there is no general decline in geyser activity." They were convinced that the great irregularity manifested by many geysers was not seasonal. As for smaller variations, they state that these "irregularities could only be detected by the aid of elaborate records, but if the quiet intervals of the Daisy Geyser are lengthened perceptibly by strong summer winds, they should be much further prolonged in the extreme cold of Yellowstone winters" (1935, p. 247). No such "elaborate records" were available for comparative data.

That Allen and Day consider atmospheric temperature as one of the factors to be considered in the gains and losses of heat in geysers is indicated by the following: "When the geyser is considered as a system subject to gains and losses of heat, we see as the factors obviously affecting these two vital conditions: (1) cooling at the surface of the water column; (2) rate of influx of the cooler ground water; (3) rate of flow of the magmatic steam, or more directly, the flow of water heated by the steam" (Allen and Day, 1935, p. 247). The studies that had been made of winter and summer temperatures led Allen and Day to conclude that surface cooling in retarding geyser activity "was of a small order of magnitude."

In connection with the 1951-1952 winter observations, the author kept a monthly record of the temperatures of 138 springs in the Upper Basin. This temperature record was started in May 1951 and continued through April 1952. The temperatures were obtained at a depth of about 12 inches, and by immersing the thermometer in that portion of the spring where there was rising convection. A careful study of this record does not reveal any significant relationships between the thermal intensity and the time of the year.

While such a temperature record as the above is important in any study of seasonal influences on thermal activity, an equally great value results from the comparative data furnished for determining the waxing or waning of thermal intensity over the years. Comparisons of the 1951-1952 recordings with those of 1925-1930, shown in the following table, are interesting. The 1925-1930 temperatures given represent averages of monthly readings, and cover the season from November through April. For about half of the listed springs in the 1925-1930 studies, temperatures were determined only during the 1928-1929 winter. Of the 138 springs whose temperatures were determined by the author, only those for which comparative readings are available are included in table 1. The temperatures were taken monthly and represent averages for the November through April period.

The table shows comparative temperatures for 60 springs. These comparisons do not lend support to the pessimism of a declining geyser activity that permeates much of the literature on the hot springs of Yellowstone Park (Allen and Day, 1935, p. 188). Forty-eight of the springs show an average gain in temperature of 2.5°. The average loss for the 12 springs which show a decline in temperature is 4.8°. The average increase for all springs in the study is 1.3°. It is significant that the great majority of the eruptive springs show an increase in temperature. This is in harmony with statistical data that there has been an increase in geyser activity since the 1946 season over that of the preceding 25 years.

The 4.8° decline in temperature of the Morning Glory Pool does not necessarily reflect an overall decline in thermal energy, and the same would be true for other springs. A study of the hot springs discloses that the temperatures of many are cyclic, periods of increase which are followed by periods of waning energy. A decrease in temperature for the Morning Glory was noted in the late thirties, lasting until 1943. From 1944 until 1951 its temperature was equal to or in excess of the 77.3 average shown for 1925-1930. Since June 1951 there has been a decline in temperature. Many nonerupting springs, like the geysers, show great irregularities in temperature. In the case of the geysers, the periods of lessened temperature are reflected by infrequent activity or dormancy.

For the geysers showing a decrease in temperature over that of the 1925-1930 record, the decline in the Sawmill and Vault no doubt results from the fact that some temperatures were taken during 1925-1930 at the time of eruptions or in the early recovery stages following eruptions. Temperatures taken by the author were at the time of a quiet phase, the temperature at this time being less than at the time of an eruption or immediately thereafter. The Sawmill and Vault are as active now as they were when the 1925-1930

TABLE 1

Comparative Temperatures

	1925-1930 Average	1951-1952 Average	Gain or Loss
Morning Glory	77.3°C.	72.5°C.	−4.8°C.
Sentinel	93.7	94.4	+0.7
Mortar	92.8	94.4	+1.6
Fan	92.3	94.1	+1.8
Spiteful	92.1	93.2	+1.1
Riverside	93.9	94.3	+0.4
Spa	82.6	87.0	+4.4
Rocket	93.2	93.8	+0.6
Grotto	93.9	94.4	+0.5
Giant	94.4	96.4	+2.0
Mastiff	93.8	96.1	+2.3
Oblong	91.9	93.2	+1.3
Inkwell	91.7	94.4	+2.7
Daisy	87.9	91.1	+3.2
Brilliant	87.8	91.0	+3.2
Comet	92.7	94.5	+1.8
Splendid	91.9	94.5	+2.6
Punch Bowl	93.7	95.0	+1.3
Black Sand Pool	92.2	92.4	+0.2
Whistle	64.2	94.3	+30.1
Spouter	92.8	92.9	+0.1
Green Spring	75.7	80.5	+4.8
Emerald	67.8	62.8	−5.0
Castle	93.3	93.7	+0.4
Tortoise Shell	94.0	95.6	+1.6
Crested Pool	91.4	93.6	+2.2
Witches Cauldron	93.1	94.4	+1.3
Calida	82.6	80.0	−2.6
Chromatic	82.5	77.7	−4.8
Beauty	75.5	67.0	−8.5
Wave	60.5	63.1	+2.6
Economic	80.1	71.0	−9.1
Turban	90.0	92.8	+2.8
Grand	77.7	80.0	+2.3
Bulger	91.4	93.3	+1.9
Sawmill	89.5	83.0	−6.5
Churn	93.8	80.0	−13.8
Tardy	88.0	91.1	+3.1
Spasmodic	92.0	93.7	+1.7
Lion	93.5	94.5	+1.0
Lioness	94.5	94.9	+0.4
Big Cub	94.0	95.2	+1.2
Little Cub	92.2	94.4	+2.2
Ear	93.8	94.0	+0.2
Beach	86.1	88.2	+2.1
Doublet	80.2	89.7	+9.5
Sponge	94.5	95.4	+0.9
Teakettle	93.4	93.7	+0.3
Giantess	93.6	94.4	+0.8
Vault	87.6	86.1	−1.5
Infant	93.3	93.4	+0.1
Beehive	93.7	95.2	+1.5
Cascade	87.6	86.6	−1.0
Blue Star	86.2	86.3	+0.1
Chinaman	93.5	93.7	+0.2
Old Faithful	94.5	94.4	−0.1
Artemisia	79.9	84.3	+4.4
Gem	77.3	85.9	+8.6
Sapphire	94.0	95.5	+1.5
Silver Globe	91.8	91.7	−0.1

temperatures were taken. The Economic is dormant, so a decline in temperature would be expected.

While the monthly temperature determinations for the 1951-1952 period do not reflect any relationship between thermal energy and the season of the year, it does not necessarily follow from this that the water in open craters is unaffected by the cold blasts of winter. As has been stated, in recording temperatures during the 1951-1952 study, the thermometer was placed in that section of the spring where there was definite upwelling of water. Temperatures here, as the records show, proved to be relatively constant when taken under similar functional conditions. This was not true at the outer margins of many of the springs. That much cooler water was being carried into the deeper recesses in winter than in summer was apparent at a number of springs, particularly the Daisy and the Grand. In view of the general belief that surface cooling delays the Daisy its case will be considered under the following heading.

HOW SUBTERRANEAN CONNECTIONS WITH OTHER SPRINGS MIGHT AFFECT GEYSER BEHAVIOR

As stated above, the Daisy is known to have underground connections with other springs and geysers. These connections have occasionally been responsible for great irregularity in the Daisy, and at times, dormancy. From late in the last century, until the 1951 season, its eruptive function suffered little from the diversion of any of its thermal energy to other connecting units. Its eruption intervals varied from about 75 to 120 minutes. For weeks at a time there would be little deviation from the average, except on those days when strong winds blew over its surface. For many years it has been observed that a wind of near 20 miles' velocity or over increases the length of the Daisy's quiet phase, and the stronger the wind the longer the interval. Winds of 30 miles' velocity will increase the average eruption interval by at least 20 to 30 minutes. Such observations as this led to the theory that the increased evaporation resulting from the wind cooled the water in the bowl of the Daisy, thereby requiring a longer recovery period between eruptions. Apparently for want of a better theory, the above explanation was the only one given on why winds delayed the Daisy.

That factors other than surface cooling resulting from the wind might be responsible for delaying the eruption time of the Daisy first became apparent to the author during the 1938 season. During that season Bonita and Brilliant Pools and Splendid Geyser, first one and then another, would periodically overflow. Normally the water in these units stands at the upper rims of their craters, except at the time of Daisy's or Splendid's activity, when a pronounced drop in water level in all craters takes place. It was observed that an overflow from any of the units connected with the Daisy reduced it to a dormant status as long as the overflow lasted. Proof that it was the overflow, say, from Bonita Pool (Marler, 1951, p. 335), that was responsible for the Daisy's inactivity was found by damming Bonita's overflow channel with fragmental sinter. Hydrostatic pressure was so sensitive that immediately the thermal energy directed away from Daisy to Bonita was channeled back.

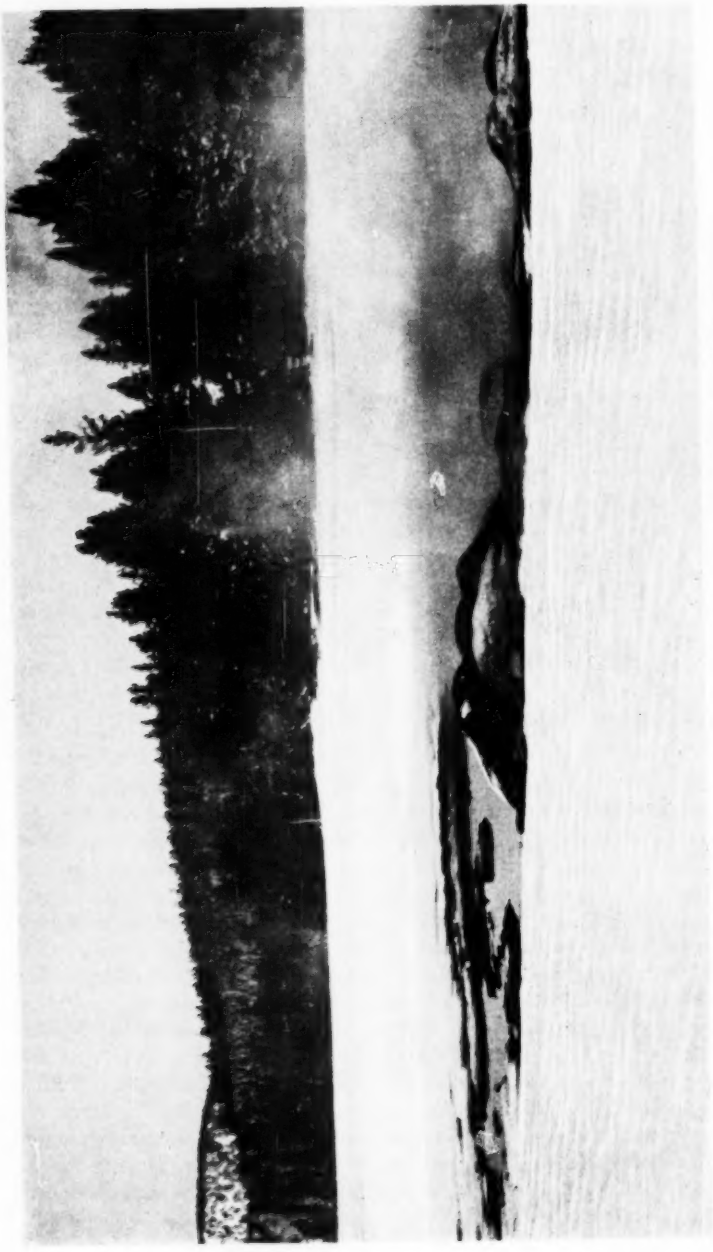
and in a short time Daisy responded by erupting. This experiment was repeated a number of times that season, and has been many times since, not only on Bonita but also on Splendid and Brilliant. Without exception the Daisy's response has been to resume normal activity.

These experiments in the Daisy Group suggested that the delayed activity of the Daisy, occasioned by the winds, was not because of surface cooling but because wind blowing over its wide collecting basin served as an agent in increasing the rate of overflow from its crater. Without a wind, water will normally overflow the sides of the crater from 20 to 30 minutes before an eruption. Heavy winds will more than double this time of overflow. This is accomplished by increasing the rate of discharge, thereby dissipating the thermal energy, which in turn increases the length of the interval. The experiments with Daisy's connecting units revealed its high degree of sensitiveness to any loss of water. The winter observations offered further confirmation to the contention that it is loss of water and not loss in surface temperature that delays the Daisy on windy days.

Following the rejuvenation of the Splendid Geyser on December 24, 1951, the Daisy was largely dormant for the remainder of the winter. There was a 3-day recovery period in February. Some of the recovery periods lasted for only one eruption, the Splendid then again took over. During the periods of dormancy in the Daisy the water would rise in its basin to just about the overflow stage, where the level would stay constant until the Splendid erupted or the Daisy again became active. During some of Daisy's dormant periods, when arctic air conditions prevailed, the water in the outer margins of its collecting basin froze solidly (pl. 1). That such a situation would greatly intensify surface cooling over evaporation effects from summer winds is patent. Yet it was during such a condition of icing that the thermal energy was observed on two occasions to shift from the Splendid to the Daisy, with the Daisy erupting on a period of overflow no longer than is average in summer.

The fact that geyser activity in open-crater geysers is as intense during subzero weather as during summer eliminates surface cooling from winds as the true explanation for Daisy's delayed activity on windy days in summer. That the Daisy is the only geyser in the Upper Basin whose function has been suspected of being affected by the winds would of itself make such an explanation questionable. The winter observations, as well as the experimental work on the units connected with the Daisy, are convincing that the delayed activity resulting from the wind is caused by the increased rate of discharge. The wind literally rolls a heavier flow of water over the side of Daisy's collecting basin, thereby dissipating its energy. Only one other geyser in Yellowstone Park is known to be affected by the wind—the Morning Geyser in the Lower Basin. It is even more sensitive to wind than the Daisy. The author has no record that the Morning has ever erupted when winds were strong enough to produce a wave action over its large collecting basin. These waves increase the rate of discharge, upsetting a delicate balance. During the 10 days of subzero weather in November 1952, the air was calm, and the Morning's eruption schedule was unaffected.

PLATE 1



During the Daisy's dormant periods, when arctic air conditions prevailed, the water in the outer margins of its shallow collecting basin froze solidly.

REASONS FOR SEASONAL CONSTANCY IN THERMAL ENERGY

A number of theories have been proposed to account for the seeming seasonal constancy in the thermal energy for the hot springs in Yellowstone Park. Allen and Day have given this problem considerable thought, and state: "If surface cooling is important as regards geyser intervals, it is possible that its effects may be modified by variations in the volume of ground water. In the climate of Yellowstone Park, much of the precipitation which in warmer weather finds its way into the ground is withheld from circulation in winter in the form of ice and snow. Reduction from this cause may conceivably go so far that considerably less heat is required to raise the remaining water to boiling, leaving another portion of heat supply free to compensate for the increased losses due to the winter cold. Such variations in ground water may also explain why in winter there is no general decline in thermal activity" (Allen and Day, 1935, p. 227).

If such an explanation as the above should prove valid, Allen and Day admit that it would demand that "changes in volume of ground water must be attended by similar variations in discharge." Winter measurements of discharge have never been attempted. The hot spring activity of winter, however, furnishes some very useful and forceful data. With the exception of the Lion Geyser, all the geysers in the Upper Basin for which records were kept showed an increase in discharge during the 1951-1952 winter over their previous summer record. For some geysers, the Grand, Giant, and Beehive, this increase in discharge, as reflected by their increased activity, was marked. The flow from no spring showed any evidence of diminishing. It is appreciated that such an observation as the latter is subjective, but waning discharge of any spring is immediately reflected by its effects on the algae in the drainage channels. In only one spring was any decrease of flow observed during the winter. This decrease was *not* due to changes in the volume of ground water, but to exchange of function. The Chromatic Pool ceased to overflow in February, and this cessation was immediately attended by an overflow of Beauty Pool. There is periodic exchange of function between these two springs. The shift back to the Chromatic took place in July 1952.

In explaining the absence of any marked seasonal variations in the thermal intensity in the main geyser basins, one important lead has not been given due consideration—the character of the ground through which the springs issue. Allen and Day recognized the great amount of glacial drift that is in evidence in the geyser basins, and that "the prevalence of these gravels in so many thermal districts of the Park indicates the probability that they may underlie large parts, if not the whole of this area." Yet, certain interpretations (Allen and Day, 1935, p. 123, 236) they made from their analyses of hot spring waters, which could have been better explained had they recognized an exchange of function between neighboring springs, prevented them from recognizing the importance of glacial gravels in explaining the seasonal constancy of hot spring activity.

The bore hole put down by the Geophysical Laboratory in 1929 behind the Old Faithful Inn went through 220 feet of gravel. A careful study of the mounds of most of the large cone geysers, Old Faithful, Castle, Giant, etc.,

reveals that these mounds are mere shields resting upon the glacial drift. The author's observations during the past several years have established that the majority of the hot springs in the Upper Geyser Basin are connected subterraneously, connections which appear logical when it is considered that glacial gravels *in depth* underlie the greater portion of the basin. That the gravels are much more permeable than the rhyolite is of utmost importance in accounting for the circulation and supply of water in this basin; the same is true for the Midway and Lower Geyser Basins.

It is the author's opinion that the major factor in accounting for the seasonal constancy in thermal energy is to be found in the huge storage basins afforded by the glacial gravels. Variations in seasonal precipitation, or the fact that snow and ice retard the replenishment of water during winter, represent such a small fraction of the total ground water that neither shows its effect—that is, from season to season.

POSSIBILITY OF INCREASE IN FUNCTION IN WINTER

The great increase in activity of most of the geysers during the 1951-1952 winter over their previous summer functioning seemed to lend credence to the possibility that, instead of a lessening of thermal manifestations during the winter, there was an increase in their intensity. No such general increase characterizes the average winter, however, as was made evident by the fact that the Giant, Riverside, and Grand not only continued on through the 1952 season playing on shorter intervals but through the summer showed additional decrease in average time. The increase in function during the winter over the previous summer, if it proves anything for most geysers, shows how independent they are of the season of the year. The increased activity of the Grand and the Giant are readily explicable when the behavior of other eruptive springs with which they have underground connections is taken into consideration.

The cases of the Riverside and the Beehive are not so easily understood as are those of the Giant and Grand. The author has observed for a number of seasons that in October the Riverside played on a shorter interval than during the previous summer. The activity of the Beehive during the 1951-1952 winter, and starting again in the winter of 1952-1953, was so far in excess of the activity in any recent summer season as to suggest that it, as well as the Riverside, erupts more frequently in winter than in summer. Only several winters of continuous observation could establish whether or not this is the case.

Should it be proven that some geysers play more frequently in winter than in summer, it would be logical to assume that for the basin as a whole there should be a general increase in thermal energy in winter over summer. The fact that there is no movement of ground water from the surface downward during most of the winter suggests the possibility of increased temperatures for circulating ground water.

CONCLUSION

The latitude and altitude of the geyser basins of Yellowstone Park subject them to one of the most rigorous climatological conditions in the United States, so far as temperature is concerned. The mean annual temperature determined by the U. S. Weather Bureau in the southeastern section of the park, where the main geyser basins are located, is 35°. The mean temperature for the 6-month period from April through October is 48.9°, while for the remaining 6 months it is 18.8°. The winter shows a departure of 31.1° from the summer mean. This great drop in temperature during the winter would be presumed not only to affect the surface temperature of all springs possessing large open craters but also to have a retarding effect on the function of many of the eruptive springs belonging to the fountain classification. This latter postulation becomes more tenable when consideration is given to the trigger balance which many of the geysers possess.

Some of the geysers, such as the Beehive, heat up to a state where the gains and losses in heat seem to be evenly balanced. That they remain in this state for long periods is evidenced by the observation that a small quantity of detergent will induce an immediate eruption, with no preceding eruption having taken place for weeks. Geysers so delicately balanced between gains and losses in temperature would be presumed to be sensitive to marked changes in atmospheric temperature. That such is not true is shown by the Beehive's winter activity.

The winter studies carried on to date have not succeeded in finding significant differences between the thermal intensity of summer and winter which could be explained as resulting from surface cooling. This lends support to the thesis that the cooling effects of the atmosphere on any spring, or any factor such as wind, which might produce a cooling of the surface water, leave the thermal intensity of that spring unchanged. If results of the investigations to date may be taken as a true indication, the evidence seems compelling that the frequency of a geyser's activity is not adversely affected by atmospheric temperature.

In explaining the seeming immunity of the hot springs to seasonal changes in their energy output, a number of possibilities present themselves: (1) The storage basins from which the hot springs draw their water supply are of such a voluminous character as to leave their discharge unaffected from season to season[1]. (2) The almost complete stoppage during the winter months of surface seepage in replenishing the water supply might result in the circulation of warmer ground water during that season. (3) The underground structure of at least the eruptive hot springs—observation has proven that the majority possess geyser potential—is no doubt of a character that prevents downward convection from reaching that section of the geyser's plumbing where the eruption is initiated. If the above suggestions present a true picture, they pose the chance of greater thermal intensity in winter than in summer. The behavior of the Beehive and Riverside is suggestive. Much additional data are necessary before this intriguing possibility can be answered with satisfaction.

[1] Reference is made only to those springs reposing on the main floors of the geyser basins. Springs on hillsides and glacial shoulders are affected by the season.

ACKNOWLEDGMENTS

The author wishes to express his appreciation to Superintendent Edmund B. Rogers and Chief Park Naturalist David Condon, Yellowstone National Park, and to Dr. L. C. Graton, Harvard University, for their interest and assistance in making possible a study of the thermal intensity in the Upper Geyser Basin of Yellowstone Park during the winter of 1951-1952. Appreciation is also expressed to Dr. Carl P. Swartzlow, Regional Naturalist, and to Mr. Conrad L. Wirth, Director, National Park Service, for their interest in the publication of the report on the winter observations.

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Yellowstone National Park
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[American Journal of Science, Vol. 252, January 1954, Pp. 55-59]

DISCUSSION

ROLE OF FROST THRUSTING IN THE FORMATION OF TUSSOCKS*

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Recently we published a discussion of cottongrass tussocks, peat rings, vegetation polygons, characteristic plant forms, and vegetation patterns of tundra areas in northwestern Alaska (Hopkins and Sigafoos, 1951). These forms and patterns result largely from interaction of frost processes and living plants. From their structures we conclude that frost thrusting—lateral soil movement during freezing—plays a major role in their development. Tussocks of cottongrass (*Eriophorum vaginatum Linnaeus* ssp. *spissum* (Fernald) Hultén) in particular owe their height in part to differential upward movement of soil beneath the tussocks as a result of lateral soil movements in and beneath freezing mineral soil in the intertussock areas, according to our hypothesis (Hopkins and Sigafoos, 1951; Sigafoos and Hopkins, 1951).

Taber (1952) recently stated that our discussion does not conform to present knowledge of the mechanics of frost action. He criticizes our reliance upon frost thrusting as a major factor in frost action and objects especially to our hypothesis that frost thrusting contributes to the development of tussocks. He suggests instead that tussocks are heaved out of the ground in a manner similar to the heaving of fence posts or boulders. Finally, he objects to our ascription of various frost phenomena to expansion in volume during freezing.

Like many investigators of microrelief features formed by frost action in soil, we have had little experience in experimental soil mechanics, and therefore have had to rely upon the thorough and thoughtful work of Taber, Beskow, and others. Most laboratory studies, however, have been concerned with small-scale frost action and have been conducted in carefully controlled systems in which only one or two factors are allowed to vary. Most experimental work in soil mechanics, moreover, has dealt with inert, nonorganic materials such as mineral soils and has not been concerned with plants. An understanding of the controlled experiments is essential to a proper understanding of geomorphic processes in nature; but when applying laboratory results to field problems one is dealing with systems that are much more complex and variable and therefore are not strictly analogous.

We recognize with Taber and other investigators that volume increase[1] in soils that freeze in the presence of abundant moisture results from the

* Publication authorized by the Director, U. S. Geological Survey.

[1] Taber's objection to our discussion of "expansion of volume" in freezing soils appears to be largely a problem in semantics. The mechanics of the volume increase discussed here and outlined in our previous paper (Hopkins and Sigafoos, 1951, p. 61) were first described in Taber's classical papers (1929, 1930). Taber prefers to speak of this volume increase as "frost-heaving," but because significant lateral expansion also seems to be involved, we prefer a more noncommittal term.

segregation of lenses of clear ice composed largely of vertical crystals. We agree, further, that this volume increase is relieved chiefly by frost heaving or upward displacement of the surface. The internal structure of many frost-formed microrelief features leaves little room for doubt, however, that during their development material also is displaced forcefully in horizontal directions. The evidence includes the tongues of silt injected into or beneath the marginal ridges of peat or tussock rings and tussock-birch-heath polygons (Hopkins and Sigafoos, 1951, fig. 25, 30) and the well-known vertical and tangential orientation of platy rocks at the margins of the fine-grained centers of stone polygons and strips (Lundqvist, 1949, p. 335; Sharp, 1942, p. 274).[2] The forceful lateral displacement that produced these structures we have termed "frost thrusting."

More than one mechanism probably is involved in frost thrusting. Stress distributions in the fine-grained mineral-soil centers of stone or vegetation rings, polygons, and stripes resemble those in a system open in regard to expansive forces only during the early winter stages of freezing. After freezing temperatures have prevailed for a short period, bare mineral soil or soil covered with a thin layer of peat freezes to depths of a few inches and the frozen layer constitutes a rigid, arched carapace that is not easily deformed. This carapace is anchored in stones or peat at the margins of the structures so that the center is, in Taber's words (1952, p. 713), bridged. Thick layers of peat and tussocks of cottongrass do not freeze, or freeze to shallower depths, thus forming weak places in the rigid carapace that are essentially open to upward movement of fluid soil. The carapace formed in the intertussock areas of polygons is especially rigid because it consists of a layer of fibrous peat interwoven with woody stems and roots frozen to mineral soil and firmly anchored to the marginal peat. As freezing continues, the growth of additional ice lenses at the base of the frozen layer exerts pressure that cannot be relieved by additional upwarping of the rigid carapace. Consequently, part or all of the pressure developed by the growing crystals is relieved downward by lateral squeezing out of soil within the still unfrozen silt layer. Tongues of this unfrozen silt are injected into peripheral unfrozen peat, laterally and upward beneath tussocks, or laterally against canted tabular stones that then are forced into a vertical position. Lateral thrusting of this type was demonstrated by Taber (1929, p. 455) in an experiment in which columns of sand and clay in contact with one another were frozen in the presence of a supply of water. Unfrozen clay was thrust into adjoining unfrozen sand by pressure developed by ice lenses growing in the upper part of the clay. Washburn (1950, p. 5) considered forceful displacement of unfrozen soil resulting from hydrostatic (cryostatic) pressure developed in overlying freezing layers to be the principal mechanism involved in the development of some frost-formed ground patterns.

It is probable, we believe, that significant horizontal pressures also are developed within freezing soil itself after frost has penetrated to depths of a

[2] Although we agree with Taber's explanation (1943, p. 1433) of the migration of stones across the surface of the fine-grained centers of stone polygons, the vertical orientation of platy stones at the margins also required the operation of significant lateral forces within the fine-grained centers.

few inches. Beskow (1935, p. 30-31) shows that the orientation of lenses of clear ice is determined by the structure of the soil as well as by the shape of the soil surface. In massive, structureless material, ice lenses are oriented parallel to the cooling surface, and the ice crystals within the lenses are oriented normal to the cooling surface. The orientation of the lenses also is influenced, however, by the orientation of physical discontinuities such as stratification (Beskow, 1935, fig. 16) or surfaces of pebbles. Root casts, contorted peat fragments, and vertical desiccation cracks resulting from withdrawal of moisture to freezing layers at higher levels (Taber, 1929, p. 450-452) furnish other potential fractures in which ice crystals can grow and exert laterally directed pressures. Beskow also emphasizes that pressure affects the orientation of ice crystals in freezing soils, but "it is essentially the pressure that occurs due to increase in volume during freezing and not the preexisting pressure in the soil" (Beskow, 1935, p. 30).

These concepts are contrary to Taber's recent generalized statement (1952, p. 713) that "when fine-grained soils freeze slowly, water segregates to form layers or lenses of ice consisting of prismatic crystals oriented normal to the surface" and that the weight of the overlying soil determines the pressure developed by growing ice crystals. The concepts do agree, however, with Taber's earlier demonstrations of the development of ice crystals in vertical desiccation cracks in clay and silt (Taber, 1929) and his demonstration that ice crystals grow outward from large particles in soil consisting of sand and clay (Taber, 1929, p. 439). Ice crystals growing in such vertical and inclined fractures can develop pressures in a confined soil mass that may be adequate to account for many of the observed evidences of frost thrusting.

It is difficult to evaluate the relative importance of lateral thrusting within freezing soils on the one hand and lateral, hydrostatic displacement of fluid soil beneath the freezing layer on the other, in producing observed thrust structures in various frost-formed microrelief features. Our observations have ceased after frost has penetrated a few inches into the soil; thus observations are lacking of the orientations of ice lenses and crystals at the critical season, when frost has penetrated deeper. In our previous discussion of the origin of tussocks, peat rings, and other features (Hopkins and Sigafoos, 1951, p. 75), we emphasized the mechanism of thrusting within the freezing layer and probably gave inadequate consideration to the importance of hydrostatic squeezing of deeper layers. We now believe that this squeezing is the predominant mechanism in the development of frost-heaved tussocks and agree with Washburn (1956, p. 30-35) that it may play a major role in the development of polygonal, circular, and striped frost features.

The cottongrass tussocks that we described in detail (Hopkins and Sigafoos, 1951, p. 70-76) can be formed only by differential upward movement of soil beneath the tussock. We cannot accept Taber's hypothesis that tussocks are heaved like fence posts and stones by being dragged upward during freezing as the entire surface heaves, leaving voids at the base that are filled by adjacent soil when it thaws. A tussock is a single plant that grows on a small mound of mineral soil and is not an inert object to be equated with a stone or a fence post. Forming a cap, concave downward over the soil mound,

are perennial rootstocks or culm bases from which grow annual roots and stems. Directly below the cap is a tightly packed column of roots and some incorporated soil; at greater depth, the roots are increasingly wide-spaced in mineral soil. The living roots—easily recognizable by their white color—are nearly straight and extend vertically downward from the rootstocks. The dark brown, dead roots of previous seasons, however, are contorted and twisted close to the rootstocks, indicating that after the roots died strong forces were exerted upward through the soil mound against the mass of roots and rootstocks. This upward force was caused largely by hydrostatic displacement of fluid soil beneath freezing layers in the intertussock areas into the unfrozen zone below and within the tussock (Hopkins and Sigafoos, 1951, p. 70-76). The mechanism is similar to the differential heaving of a freezing sand column due to the intrusion of unfrozen clay in Taber's experiment (1929, p. 455). No surface exists within a tussock below which ice crystals can exert a force that would push a tussock upward, nor have we found any evidence that significant ice lenses form within the tussock. The roots are so widely spaced and fragile at the level of the intertussock surface that they would be torn from the tussock if the tussock were dragged upward in the grip of adjoining freezing, heaving mineral soil.

Taber (1952) stated that tussocks would settle, as do boulders, if, as we believe, the soil were fluid enough to transmit hydraulic pressure. The soil mound within the tussock remains intact throughout the summer and the tussock does not settle because: 1) adhesion of the roots to the mineral soil and physical resistance of the root mass to soil movement prevent lateral or downward migration of the tussock and the soil mound except under exceptional pressures; 2) tussocks are tied to and, in part, held in place by the surrounding vegetation mat; and 3) during summer the growing tussocks remove water from the mineral soil by transpiration and water drains from the soil mound, so that the soil beneath a tussock is less fluid than the surrounding soil.

Just as the distinctive characteristics of cottongrass tussocks in relation to frost processes must be considered in explaining their formation, so should the distinctive characteristics of vegetation be considered in geomorphic studies. We believe that nearly all geomorphic processes have acted on the land surface through plant cover, a factor that is difficult to introduce into standard soil mechanics tests. It is therefore important that the complexity of natural systems be appreciated.

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COMMUNICATION

ASSOCIATION OF PERTHITIC MICROCLINE WITH STRESSED QUARTZ

In an article in this Journal (v. 250, 1952, p. 281-296) I pointed out that perthitic albite was more abundant in the microcline of calcalkaline granites whose quartz had been extensively granulated than in those in which quartz was either uncrushed or undulant. From this I argued that shearing stress accelerates, if it does not actually induce, the unmixing of soda-poor microcline, suggesting that (p. 294) "the effect may be compared to that of vigorous stirring on a slightly supersaturated salt solution. . . ." At the time I was quite unaware that this was almost a direct quotation from Harker.

In his discussion of retrograde metamorphism of regionally metamorphosed rocks (*Metamorphism*, 2d ed., p. 353), he attributes the great abundance of perthite in such rocks to the ". . . extremely gradual decline of temperature . . ." but adds that "probably any recurrence of stress-conditions during the time of cooling will promote the change, the effect being analogous to that of stirring a supersaturated liquid solution."

Independent "discoveries" of this sort must be anticipated in any field as well explored as the petrography of the feldspars, but it is rather rare to find them couched in identical language. If it were argued that my "discovery" was simply an unconscious memory of Harker's, I would have no effective counter. Instead of developing a new hypothesis of my own I seem to have been documenting a casual assertion of his. The idea, of course, was not novel with Harker, but like his predecessors he presents no evidence one way or the other. That I did devise a fairly critical test of his hypothesis may perhaps mitigate my failure to acknowledge his priority.

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ESSAY REVIEW

The Earth, Its Origin, History, and Physical Constitution, 3rd ed.; by HAROLD JEFFREYS. P. 392; 30 figs., 10 pls. Cambridge and New York, 1952 (Cambridge University Press, $13.50).—The first edition of this widely known work appeared in 1924, the second in 1929. Rapid development in knowledge and application of geophysical principles during the past quarter-century is reflected in the new edition. The size of the book is increased only moderately, since four chapters dealing with the origin of the Solar System have been omitted. The author does not consider any present theory of origin entirely satisfactory, but he states (p. 268), "The most important point for geophysical purposes is whether the Earth was formerly fluid. Most existing theories agree that it was." Keeping the word *origin* in the title therefore seems hardly justified, though the early history of the Earth receives considerable attention in the new edition.

The new book deals essentially with the physical constitution of the Earth. Starting with a brief chapter on mechanical properties of rocks, the author goes on to the theory of elastic waves, introductory to "Observational Seismology," the longest and one of the most useful sections of the book. The difficulties in problems faced by seismologists are outlined, to give an appreciation of uncertainties as well as firm values in the best results now available. To geologists the points of most immediate interest are continued confirmation of the Mohorovičić discontinuity at a few tens of kilometers depth, and indications of highly varied composition, horizontally as well as vertically, above this lower boundary of the "crust." Jeffreys' suggestions that an upper layer may be a mixture of obsidian with granitic rock, and that an intermediate layer consists chiefly of tachylyte will be greeted by petrologists with skeptical interest. Jeffreys himself notes that "other identifications would be possible," but apparently he is confident that in continental areas the upper and intermediate layers are distinct and continuous entities, a view that is questioned by some other geophysicists.

Seismic evidence on deeper zones is treated in some detail. The curve for velocity of the primary wave is shown steepening in a straight line to depth about 400 km. Thus Jeffreys is not willing to indorse Gutenberg's concept that velocity falls off in a zone between 80 and 150 km in depth. Otherwise the velocity curves favored by these two geophysicists are in general agreement —abrupt steepening at about 400 km, falling off of acceleration to 900 or 1000 km, and nearly uniform increase of velocity through the rest of the mantle, to 2900 km depth. Jeffreys favors iron as the chief component of the core, with change in phase near 5100 km depth.

The treatment of seismology is followed by five chapters dealing with geodetic matters, including relations of the Earth to the Moon. As the author states, "The determination of the Earth's axes is the main object of geodesy." He gives considerable attention to the figure of the Earth and to gravity values, and quotes the view of Barrell that isostatic compensation is not exact and that there are conspicuous residual loads implying appreciable strength in the crust. In studying regional groups of anomalies, Jeffreys takes exception to a common view that isostatic reductions are desirable, and uses instead free-air values, maintaining that systematic errors are thus avoided. His world

anomaly chart shows an alternation of positive and negative group values in low north latitudes, the positive maxima at about 10° and 170° east longitude, the negative at 90° east and 90° west longitude. The longitude terms in the solution contain uncertainties; but "if they are genuine they imply departures of the geoid from the spheroid of the order of 100 m." This estimate is much less than that of some well-known geophysicists whose interpretations of gravity values represent the figure of the Earth as definitely triaxial.

In Chapter 9, "The Age of the Earth," Jeffreys notes ten methods of approach that have led to estimates, but gives only passing notice to four of these, such as sodium in the ocean and the Sun's radiation, which now have only historical interest. On testimony from the other six he concludes, "There is general consistency with an age in the neighborhood of 2000-3000 million years." It should be noted, however, that the estimate of 2000 million based on recession of the nebulae has been greatly increased since the new edition appeared. Moreover, since lead-uranium ratios in rocks of the crust supply the most reliable figures, and the highest value now in hand from this method gives an age in excess of 2100 million years, at least the lower limit in Jeffrey's suggested brackets may well be moved up.

Jeffreys' view that our planet passed through a fluid stage has remained unshaken by recent revivals of the concept that the Earth has always been solid. He points out that free fall of cold particles into the Earth's field would supply as much as 15,000 calories per gram, and thus fairly rapid accretion of such material would provide heat for an early fluid stage. Evidence for density stratification is convincing. He rejects the Kuhn-Rittman theory that the core consists of compressed solar matter, largely hydrogen. Convection would have transferred both heat and matter to the surface; moreover consistent application of the theory would require that Jupiter and Saturn be much denser than the Earth, whereas the large planets have much lower density. All our evidence regarding the core is consistent with the concept that iron, the abundant metal in meteorites, was separated during a fluid stage. Such a stage is indicated also by concentration of the more acid rocks in a surficial shell. Those who assume an all-time solid Earth suggest addition of this outer material during a final stage of accretion, when the Earth passed through an atypical swarm of small bodies. "This seems a desperate expedient."

A fluid state in the outer part of the core, indicated by its response to seismic waves, is explained logically by temperature above the melting point for metals, blanketed by the silicate mantle with somewhat higher melting point. Jeffreys reasons that in an early stage of the history cooling by convection proceeded rapidly, that solidification of the mantle occurred quickly from the bottom upward, and that heat has continued to flow, through materials that have low conductivity, to the present time. What part of the heat flow now measured at the surface comes from the Earth's interior, and how much of it is generated by radioactive substances in the outer part of the mantle? Data now in hand indicate that all the heat would be supplied by 20 km thickness of granite with about the average content of uranium, thorium, and potassium in known granites. Jeffreys suggests that radioactivity

is confined to a thin outer shell, or that it falls off with depth so rapidly as to be effectively limited to such a shell. He cites the argument of V. M. Goldschmidt that the atoms of uranium and thorium, because they are too large to fit into any silicate lattice, must have remained in solution and been forced upward as the mantle solidified. He quotes also data cited by Holmes on increasing content of radioactive elements with decreasing age in several Finnish granites.

These and other considerations help Jeffreys to reconcile what is known about radioactivity with his concept of a cooling Earth. He goes on, in Chapter 11, to discuss "the origin of the Earth's surface features," starting with thermal contraction as a possible cause of crustal deformation. His estimate of the amount of compression available includes an important quantity overlooked earlier, when he reasoned that no appreciable cooling has occurred below a depth of a few hundred kilometers. His earlier equation assumed infinite radius of curvature, and thus in effect an infinite supply of heat. Now he realizes that heat must have flowed continuously from the core outward, and he computes a fall in temperature of about 45°C. at the center, 100° at depth 600 km, and increasingly more with shallower depth, with average about 500° down to 400 km. Assumed thermal contraction of 5 percent in this shallow zone, supplemented by loss of volume through crystallization and by expulsion of water, implies shortening of the circumference by 130 km. To this, however, may be added as much as 400 km owing to cooling in the deep interior. This contribution increases uniformly with time, and thus supplies greater energy for deformation as the outer stabilized shell grows thicker and more resistant. The present thickness of this passive shell, above a "level of no strain," is estimated as about 100 km.

Jeffreys rejects estimates of crustal shortening based on attempted restoration of deformed sedimentary beds, such as Heim's suggested ratio of 3 or 4 to 1 in the Swiss Alps. So much shortening and implied thickening in a shell 30 or 40 km thick would result in phenomenal mountain heights. On assumption of 33 km total thickness for the sialic and intermediate layers, resting on dunite, and average "primitive" height of 3.2 km in the Alps and Rocky Mountains, shortening would be in the ratio about 1.66 to 1. Geologists can follow this reasoning if *primitive height* is taken to mean *total uplift* related to the deformation. Since orogeny in any belt has been distributed through considerable time, restoration of the height lost by attendant erosion is not a simple matter. We can hardly accept without qualification Jeffrey's conviction that such erosion, by removing loads locally, has set in play general isostatic uplift which carried residual masses to exceptional height. His distrust of deformed sedimentary sections as a key to total deformation is of course shared by students of tectonics, many of whom are sympathetic to the suggestion that superficial yielding under gravitational stress has been responsible for large horizontal displacement in some mountain belts.

In Jeffrey's view, then, the contraction concept meets the test of energy requirement through geologic time. He finds the hypothesis favored also by the global pattern of Tertiary mountain belts. Notwithstanding his obvious preference, he outlines the theories of convection presented by Pekeris and

Hales, with some favorable comment but with objections to the assumption of "finite viscosity" in the mantle.

Jeffreys considers also the problem of the origin of continents and ocean basins, and favors a form of the hypothesis offered by G. F. S. Hills, which assumes segregation of crystals by convection currents in an early stage of solidification. Among other subjects that receive brief attention are shifts of the Earth's axis, continental drift, growth of the hydrosphere, lunar craters, and rift valleys. The book is required reading for students of tectonics, who will detect a few weaknesses from the geologist's point of view, but will find much that is helpful and stimulating. Those who are familiar with Jeffrey's mode of presentation will not expect light reading; even readers fairly equipped in mathematics will encounter hurdles, since some of the derivations are of necessity outlined sketchily. Some compensation is provided in the clear English of the text.

CHESTER R. LONGWELL

REVIEW

Geology of the Caballo Mountains; by VINCENT C. KELLY and CASWELL SILVER. P. 286; 19 pls., 26 figs. (9 of them, including colored geologic map, folded, in pocket). Albuquerque, 1952 (University of New Mexico Press, $2.50, paper cover; $5.00, clothbound).—The Caballo Mountains, a north-trending range about 30 miles long, are in southern New Mexico, directly east of the Rio Grande River and nearly 100 miles upstream from El Paso, Texas. Maximum relief is about 3000 feet. The Elephant Butte Dam and Reservoir are near the north end of the range, and the Caballo Reservoir, made by another dam across the Rio Grande, parallels the range on the west. The principal town of the area has the unusual name *Truth or Consequences.*

Precambrian granite and metamorphic rocks exposed near the west base of the range are overlain by a thick stratified sequence, with general inclination somewhat north of east. All Paleozoic systems are represented in a total thickness about 5000 feet. No Triassic or Jurassic rocks are present, but there is a thick Cretaceous section, starting with beds referred to the "Dakota group," the lower part of which may be of Lower Cretaceous age. Tertiary basin deposits, including the Santa Fe formation, are widespread in the low ground on both sides of the range. Dikes, sills, and remnants of basaltic lavas are dated as Tertiary and Pleistocene.

The sedimentary formations record no orogeny earlier than late Cretaceous time. Some moderate folding may be as late as Miocene, and movement on steep faults was considerable as late as Pleistocene time. Integration of the Rio Grande drainage was not completed before the Pliocene and perhaps not until early Pleistocene.

The report is well prepared, with clear maps and diagrams and an extensive bibliography. The geologic map, on the scale one inch to a mile, does not have a topographic base.

CHESTER R. LONGWELL

Publications Recently Received

Publications of the XIXe Congrès Géologique International, Alger, 1952, as follows: Symposium sur les Séries de Gondwana: Curt Teichert, editor.

Symposium sur les Gisements de Fer du Monde: F. Blondel and L. Marvier, editors. Tomes I, II, III (Atlas): Carte géologique, Algérie-Tunisie: Carte géologique, Maroc-Algérie.

Monographies régionales, 1re série, Algérie: 1, Ahnet et Mouydir, by J. Follot: 2, Massif culminant de l'Ouarsensis, by L. Calembert: 3, Monts du Mellègue, by G. Dubourdieu: 4, Socle métamorphique en Grande Kabylie, by J. Thiébaut: 5, Le massif du Bou Taleb, by J. Bertraneu: 6, L'Antécambrien de l'Ahaggar (Sahara central), by M. Lelubre: 7, Les monts du Hodna (partie orientale), by J. Glaçon: 8, La région de Colomb-Béchar, by P. Deleau: 9, La bordure sud-tellienne dans le Titteri, by J. Guillemot: 10, Le Sud-Ouest de la Petite Kabylie, by M. Durand Delga: 11, Les appareils volcaniques récents de l'Ahaggar, by P. Bordet: 12, L'Atlas saharien sud-oranais, by A. Cornet: 13, Le Pays constantinois, by P. Deleau: 14, Les chaines atlasiques et la bordure nord du Sahara. Aperçu d'ensemble, by J. Flandrin: 15, Les chaînes d'Ougarta et la Saoura, by N. Menchikoff: 16, Le Bassin néogène du Chéliff, by S. N. Repal: 17, La région de Bougie, by L. Duplan: 18, Les roches cristallines du Littoral oranais, by G. Sadran: 19, La chaîne du Djurdjura, by J. Flandrin: 20, Régions sud-telliennes et Atlas saharien, by S. N. Repal: 21, Bordure nord des Hautes Plaines dans L'Algérie occidentale, by G. Lucas: 22, Le Littoral oranais, (Mouvements verticaux et anomalies gravimétriques), by Y. Gourinard: 23, Massifs cristallins du Nord constantinois, by J. Bolfa, J. Hilley, and M. Roubault: 24, L'Atlas tellien occidental, by M. Dalloni: 25, Histoire géologique de la Province d'Alger, by L. Glangeaud: 26, Aperçu stratigraphique sur la Micropaléontologie du Crétacé, by J. Sigal: 27, Genèse des gîtes phosphatés du Sud-Est algéro-tunisien, by L. D. Visse: Hors-série, La Paléontologie des Vertébrés en Afrique du nord française: by C. Arambourg.

Monographies régionales, 2eme série, Tunisie: 1, Paléogéographie, tectonique et orogénèse de la Tunisie, by G. Castany: 2, Les gîtes métallifères tunisiens, by Ch. Gottis and P. Sainfeld: 3, Les grands problèmes d'hydrogéologie en Tunisie, by G. Castany, R. Dégallier, and Ch. Domergue: 4, Les grands barrages de la Tunisie, by Ch. Gottis and R. Strohl: 5, Le Sillon tunisien, by J. Bolze, P. F. Burollet, and G. Castany: 6, Atlas tunisien oriental et Sahel, by G. Castany: 7, Sud et Extrême-Sud tunisiens, by Ch. Domergue, E. Dumon, A. F. De Lapparent, and P. Lossel.

Monographies régionales, 3e série, Maroc: 1, Géologie des gîtes mineraux marocains, by J. Agard et al.: 2, L'Industrie minière du Maroc, by L. Eyssautier: 3, Aspects de la géomorphologie du Maroc, by J. Dresch et al.: 4, Hydrogeologie du Maroc, by R. Ambroggi et al.: 5, Géologie appliquée aux grands travaux du Maroc, fasc. 1, by P. Leveque: 6, Géologie du Maroc, fasc. 1, 1re partie by G. Choubert and J. Marcais, 2e partie by G. Choubert.

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